Arctic Report Card 2007

Tracking recent environmental changes

Atmosphere	Biology
Sea Ice	Ocean
Greenland	Land

Warming and mixed signals

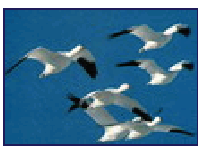
Collectively, the observations indicate that the overall warming of the Arctic system continued in 2007. There are some elements that are stabilizing or returning to climatological norms. These mixed tendencies illustrate the sensitivity and complexity of the Arctic System.

Atmosphere

Hot spot shifts toward Europe

Sea Ice

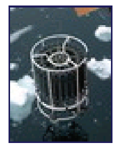
Summer extent at record minimum



Biology

• Increasing tundra shrub cover and variable treeline advance

• Up to 80% declines in some caribou herds, while goose populations double

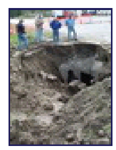


Ocean

North Pole Temperatures at depth returning to 1990s values

Greenland

Recent warm temperatures associated with net ice loss



Land

Increase in permafrost temperatures is slowing down

October 2007

Citation:

J. Richter-Menge, J. Overland, E. Hanna, M.J.J.E. Loonen, A. Proshutinsky, V. Romanovsky, D. Russell, R. Van Bogaert, R. Armstrong, L. Bengtsson, J. Box, T.V. Callaghan, M. De Dapper, B. Ebbinge,O. Grau, M. Hallinger, L.D. Hinzman, P. Huybrechts, G.J. Jia, C. Jonasson, J. Morison, S. Nghiem, N. Oberman, D. Perovich, R. Przybylak, I. Rigor, A. Shiklomanov, D. Walker, J. Walsh, and C. Zöckler (2007). Arctic Report Card 2007, http://www.arctic.noaa.gov/report07.

CAFF Conservation of Arctic Flora and Fauna

Atmosphere

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Circulation regime

The annually averaged Arctic Oscillation index (AO, a measure of the strength of circumpolar winds) was slightly positive in 2006, continuing the trend of a relatively low and fluctuating index which began in the mid-1990s (Figure A1). This follows a strong, persistent positive pattern from 1989 to 1995. The current characteristics of the AO are more consistent with the characteristics of the period from the 1950s to the 1980s, when the AO switched frequently between positive and negative phases. Initial data from 2007 shows a positive AO pattern.

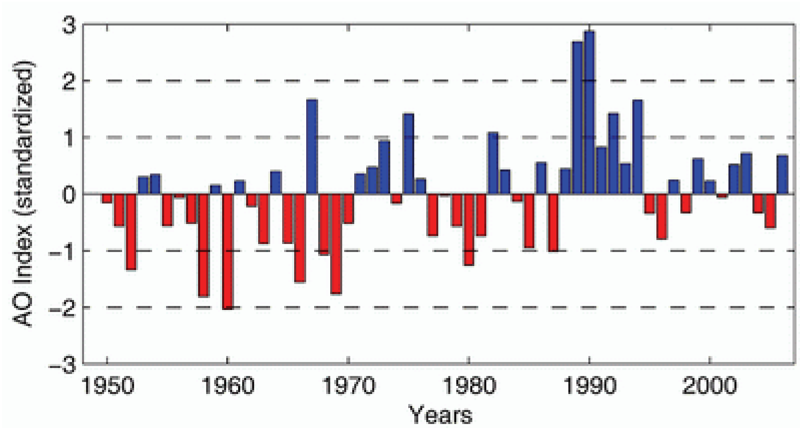

Figure A1. Time series of the annually-averaged Arctic Oscillation Index (AO) for the period 1950 - 2006 based on data from the website www.cpc.ncep.noaa.gov. (Courtesy of I. Rigor)

Surface Temperatures and Atmospheric Circulation

In 2006 the annual surface temperature over land areas north of 60° N was 1.0°C above the mean value for the 20th century (Figure A2). The surface temperature in this region has been consistently above the mean since the early 1990s. Figure A2 also shows warm temperatures in the 1930s and early 1940s, possibly suggesting a longer-term oscillation in climate. However, a detailed analysis shows different proximate causes for the 1930s compared to recent maxima. The early warm and cold periods are associated with internal variability in high-latitude circulation patterns, while the recent warm temperatures have an anthropogenic component (Johannessen et al. 2004; Wang et al 2007).

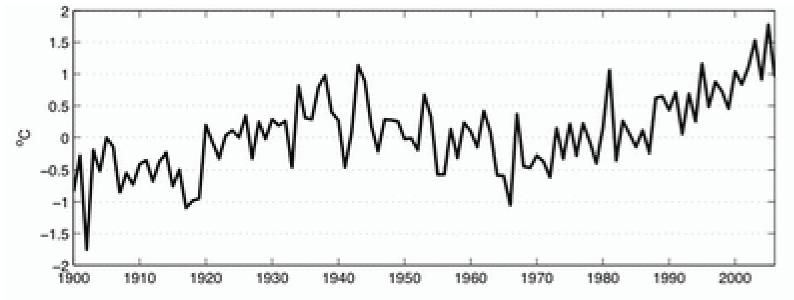

Figure A2. Arctic-wide and annual averaged surface air temperature anomalies (60° - 90° N) over land for 1900-2006 based on the CRU TEM2V monthly data set. Anomalies are relative to the 20th century average.

For winter and spring (Dec-May) in 2006 and 2007 there was an overall warm pattern (positive temperature anomalies) in the Arctic with a regional hot spot of +3-4°C near Svalbard, spreading north from the Barents Sea (Figure A3 Left). This pattern was slightly different than observed during 2000-2005 which also had the overall warm pattern, but the hot spot was closer to east Siberia. The pattern of 2006 and 2007 Dec-May sea level pressure (SLP) anomalies shows a dipole pattern with higher pressure over Asia and lower pressure over the North American side of the Arctic (Figure A3 Right). This current SLP dipole implies an anomalous northward (meridional) flow of air from the Barents Sea to the central Arctic which supports the 2006-2007 temperature hot spot through warm air advection. The 2006-2007 period continues the pattern set up during 2000-2005 with Arctic-wide positive temperature anomalies, and a meridional flow pattern toward the central Arctic. The recent 2000-2007 Arctic warm period contrasts with the two principal atmospheric circulation features of the 20th century: the Pacific North American Pattern, which was strong during 1977-1981, and the Arctic Oscillation/Northern Annular Mode/North Atlantic Oscillation, which was strong during 1989-1995 (Quadrelli and Wallace 2004, Overland and Wang, 2005). The positive phases of these two patterns gave positive temperature anomalies over the Arctic land masses, while the current pattern shows positive temperatures centralized over the Arctic Ocean. These contrasts illustrate that we are in a period of continuing uncertainty about the dominance of any one climate pattern over the Arctic.

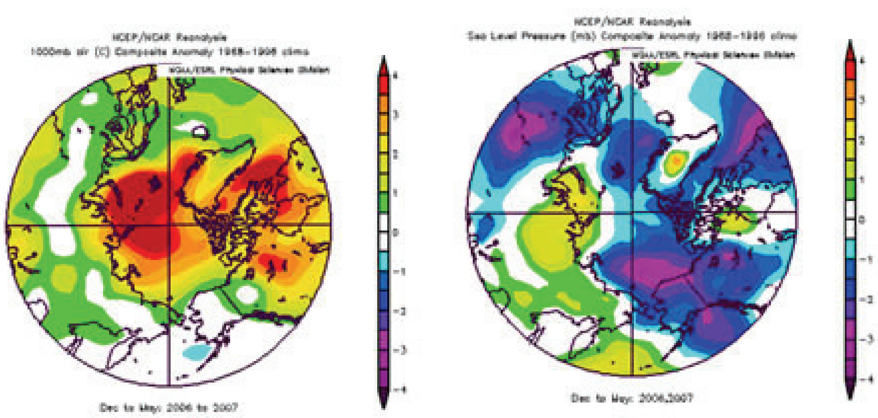

Figure A3. Left: December-May temperature anomaly composites for 2006 and 2007. Right: December-May sea level pressure anomaly composite for 2006 and 2007. All of the Arctic has positive temperature anomalies with a hot spot in the central Arctic northeast of Svalbard. The SLP anomaly pattern is a dipole, suggesting anomalous northward air flow into the central Arctic from Eurasia. The figure is based on NOAA National Centers for Environmental Prediction (NCEP) reanalysis fields via the Climate Diagnostics Center, www.cdc.noaa.gov. Anomalies are relative to a 1968-1996 base period.

End of an era for the Bering Sea?

Unlike the remainder of the Arctic, as noted above, air and ocean temperatures in the Bering Sea cooled significantly in 2006 and early 2007 compared with the previous six year period of warm temperatures (Figure A4 (ocean), Figure A3 (air)). Vertically average temperatures from an oceanographic mooring on the southeastern Bering Sea continental shelf (Stabeno et al. 2002) recorded temperatures in 2000-2005 that were 2°C warmer than earlier years, with 2005 as the warmest summer. While winter 2006 was very cold (note the drop in temperature between fall 2005 and summer 2006), the spring temperatures and ice extent in 2006 were near their climatological averages because the beginning fall 2005 temperatures were warm. Temperatures in fall 2006, in contrast, started cold and the weather pattern for November-December 2006 was also cold. The six year period of sustained of warm temperatures was sufficient to restructure the ecosystem away from Arctic conditions (Grebmeier et al. 2006). Winter-spring 2007 ended by being a relatively extensive ice year in the Bering Sea region. This suggests that it took two years for the warm ocean temperature anomalies on the Bering Sea continental shelf to dissipate. Because of this dramatic shift in ocean and ice conditions, the future state of the Bering Sea ecosystem is now less certain.

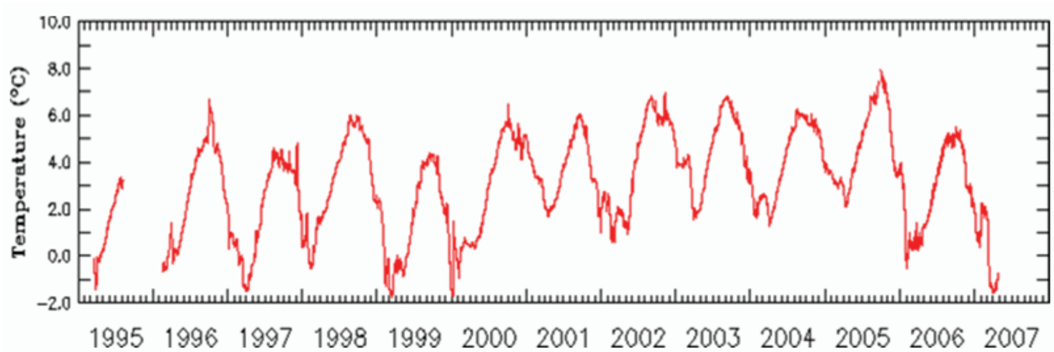

Figure A4. Ocean temperatures from a mooring on the southeastern Bering Sea continental shelf.

References

Grebmeier, J.M, and co-authors (2006) A major ecosystem shift in the northern Bering Sea. *Science, 311*, 1461-1464.

Johannessen, O.M., and co-authors (2004) Arctic climate change: Observed and modeled temperature and sea ice variability. *Tellus, 56A*, 328-341.

Overland, J.E., and M. Wang (2005) The third Arctic climate pattern: 1930s and early 2000s. *Geophys. Res. Lett., 32*, L23808, DOI:10.1029/2005GL024254.

Quadrelli, R., and J.M. Wallace (2004) A simplified linear framework for interpreting patterns of northern hemisphere wintertime climate variability. *J. Climate, 17*, 3728-3744.

Stabeno, P.J., N.B. Kachel, M. Sullivan, and T.E. Whitledge (2002) Variability of physical and chemical characteristics along the 70-m isobath of the southeast Bering Sea. *Deep-Sea Res. Pt. II, 49*, 5931-5943.

Wang M., J.E. Overland, V. Kattsov, J.E. Walsh, X. Zhang, and T. Pavlova (2007) Intrinsic versus forced variation in coupled climate model simulations over the Arctic during the 20th century. *J. Climate* (in press).

Sea Ice Cover

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Extent and Thickness

Satellite-based passive microwave images of the sea ice cover have provided a reliable tool for monitoring changes in the extent of the ice cover since 1979. During 2006 the minimum ice extent, typically observed in September, reached 5.9 million km^2 (Figure I1, bottom left panel). This marked a slight recovery from the record minimum of 5.6 million km^2 for the period 1979-2006, observed in 2005. Consistent with the past several years, the summer retreat of the ice cover was particularly pronounced along the Eurasian coastline. A unique feature was the sizeable isolated region of open water apparent in the Beaufort Sea.

The 2007 summer sea ice extent marked a new record minimum, with a dramatic reduction in area of coverage (4.3 million km^2) relative to the previous record set just 2 years ago in 2005 (Figure I1, bottom right panel). At the end of the 2007 melt season, the sea ice cover was 23 percent smaller than it was in 2005 and 39 percent below the long-term average from 1979 to 2000.

The maximum ice extent is typically observed in March. In 2006, the maximum extent was 14.4 million km^2 and set a record minimum for the ice-extent maximum for the period 1979-2006 (Figure I1, top left panel). It is notable that in March 2006 the ice extent fell within the mean contour at almost every location. In March 2007, the maximum ice extent was 14.7 million km^2 (Figure I1, top right panel).

For comparison, the mean ice extent for March and September, for the period 1979-2007, is 15.6 million km^2 and 6.7 million km^2, respectively.

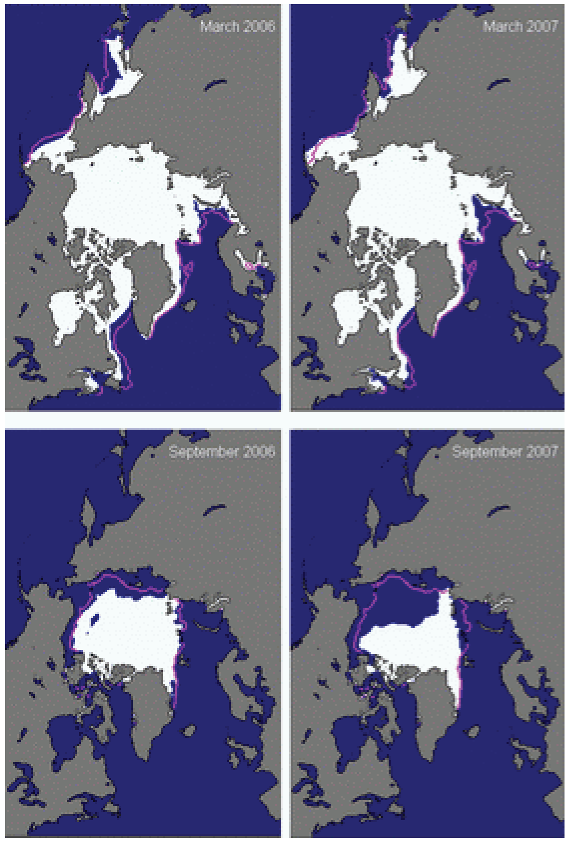

Figure I1. Sea ice extent in March and September 2006 and 2007, when the ice cover was at or near its maximum and minimum extent, respectively. The magenta line indicates the median maximum and minimum extent of the ice cover, for the period 1979-2000. The March 2006 maximum extent and the September 2007 minimum extent established new records as the lowest extents for the period 1979-2007. (Figures from the Sea Ice Index, nsidc.org/data/seaice_index)

To put the 2006 and 2007 minimum and maximum ice extent into context, the time series of the anomaly in ice extent in March and September for the period 1979-2007 is presented in Figure I2. In both cases, a negative trend is apparent with a rate of 2.8% per decade for March and 11.3% per decade for September relative to the 1979 values. The summers of 2002-2007 have marked an unprecedented series of extreme summer ice extent minima.

Ice thickness is intrinsically more difficult to monitor. With satellite-based techniques (Laxon et al., 2003; Kwok et al., 2004) only recently introduced, observations have been spatially and temporally limited. Data from submarine-based observations indicate that the ice cover at the end of the melt season thinned by an average of 1.3 m between the period 1956-1978 and the 1990s, from 3.1 m to 1.8 m (Rothrock et al., 1999). Measurements of the seasonal and coastal ice cover do not indicate any statistically significant change in thickness in recent decades (Melling et al., 2005; Haas, 2004; Polyakov et al., 2003).

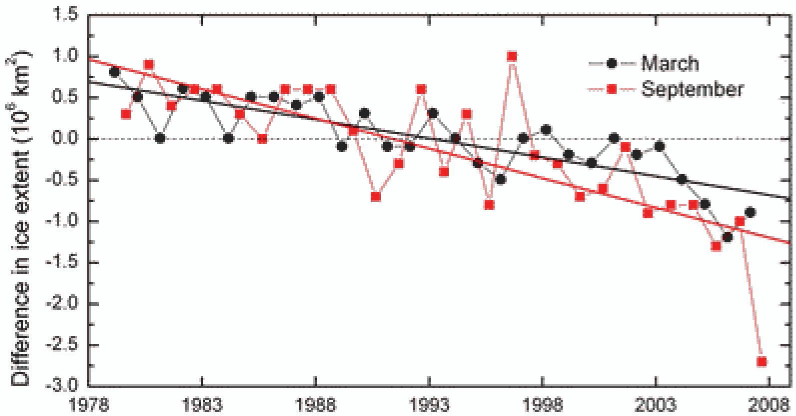

Figure I2. Time series of the difference in ice extent in March (the month of ice-extent maximum) and September (the month of ice-extent minimum) from the mean values for the time period 1979-2007. Based on a least squares linear regression, the rate of decrease for the March and September ice extents was 2.8% per decade and 11.3% per decade, respectively.

Perennial and Seasonal Ice

The Arctic sea ice cover is composed of perennial ice (the ice that survives year round, generally located towards the center of the Arctic basin) and seasonal ice (the ice around the periphery of the Arctic basin that melts during the summer). Consistent with the diminishing trends in the extent and thickness of the cover is the observation of a significant loss of the older, thicker perennial ice in the Arctic (Figure I3). Results from a simulation using drifting buoy data and satellite-derived ice concentration data to estimate the age distribution of ice in the Arctic Basin (Rigor and Wallace, 2004) indicate that the March ice cover has experienced a significant decline in the relative amount of perennial ice over the period 1958-2006, from approximately 5.5 million km^2 to 3.0 million km^2. While there is significant interannual variability, a generally downward trend in the amount of perennial ice begins in the early 1970s. This trend appears to coincide with a general increase in the Arctic-wide, annually averaged surface air temperature, which also begins around 1970 (Figure A2).

Results from a new technique employing data acquired by the U.S. National Aeronautics and Space Administration (NASA) SeaWinds scatterometer on board the QuikSCAT satellite (QSCAT) have recently become available (Nghiem et al. 2005; Nghiem et al.; 2006, Nghiem and Neumann, 2007). In the half decade of overlap with the buoy-derived results, which presently begins in 2002 and represents the period of data reprocessed to date by the QSCAT project, the two products provide consistent estimates of perennial ice in March and suggest a precipitous decrease in the perennial ice extent in the last few years.

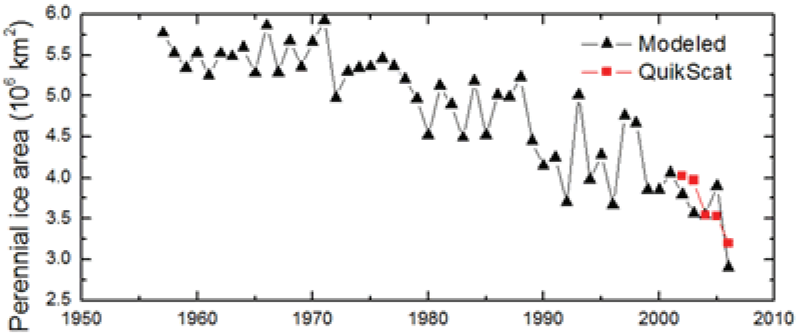

Figure I3. Time-series of the area of perennial sea ice extent in March estimated by a drift age model and satellite-derived ice concentration data and observed by the QuikSCAT scatterometer within the drift age model domain.

Figure I4 presents a comparison of the ice distribution derived from the drift age model and observed by QSCAT in March 2006. The two products provide similar results. Both indicate that the older, thicker ice is concentrated in the western Arctic basin. This result is consistent with the dominant ice circulation patterns in the Arctic (see Figure O1). Ice residence times are typically longer in the western Arctic in the region of the Beaufort Gyre. The eastern Arctic is dominated by the Trans Polar Drift, which carries sea ice out of the Arctic Basin via the Fram Strait.

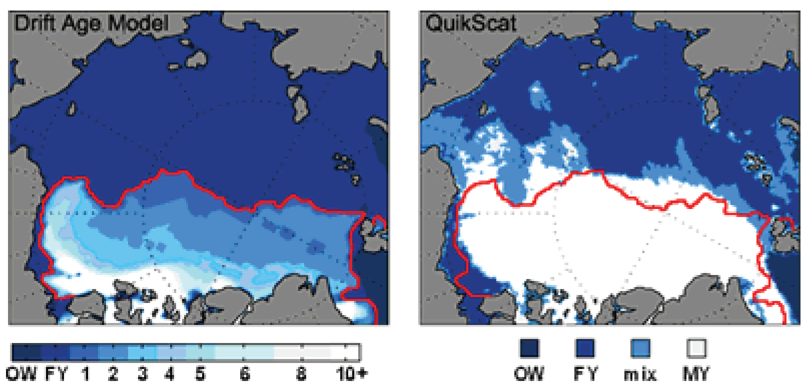

Figure I4. Comparison of sea ice distribution estimated using the drift-age model (March average, left panel) with QSCAT observations (21 March 2006, right panel). The red line in both panels indicates ice age older than 1 year (i.e. perennial ice) as estimated by the drift age model.

The development of a relatively younger, thinner ice cover coincided with a strong, persistent positive pattern in the AO from 1989 to 1995 (see Figure A1). These characteristics make the current ice cover intrinsically more susceptible to the effects of atmospheric and oceanic forcing. It is of crucial importance to observe whether the sea ice cover will continue its decline or recover under the recent more neutral AO conditions (Lindsay and Zhang, 2005).

References

Haas, C. (2004) Late-summer sea ice thickness variability in the Arctic Transpolar Drift 1991-2001 derived from ground-based electromagnetic sounding. *Geophys. Res. Lett., 31*, L09402, doi: 10.1029/2003GL019394.

Kwok, R., H.J. Zwally, and D. Yi (2004) ICESat observations of Arctic sea ice: A first look. *Geophys. Res. Lett., 31*, L16401, doi: 10.1029/2004GL020309.

Laxon, S., N. Peacock, and D. Smith (2003) High interannual variability of sea ice thickness in the Arctic Region. *Nature, 425*, 947-950.

Lindsay, R.W., and J. Zhang (2005) The Thinning of Arctic Sea Ice, 1988-2003: Have We Passed a Tipping Point? *J. Climate, 18*, 4879-4894.

Melling, H., D.A. Riedel, and Z. Gedalof (2005) Trends in the draft and extent of seasonal pack ice, Canadian Beaufort Sea. *Geophys. Res. Lett., 32*, L24501, doi:10.1029/2005GL024483.

Nghiem, S.V., M.L. Van Woert, and G. Neumann (2005) Rapid formation of a sea ice barrier east of Svalbard. *J. Geophys. Res., 110*, doi:10.1029/2004JC002654.

Nghiem, S.V., Y. Chao, G. Neumann, P. Li, D. K. Perovich, T. Street, and P. Clemente-Colon (2006) Depletion of perennial sea ice in the East Arctic Ocean. *Geophys. Res. Lett., 33*, L17501, doi:10.1029/2006GL027198.

Nghiem, S.V., and G. Neumann, Arctic Sea-Ice Monitoring (2007) *2007 McGraw-Hill Yearbook of Science and Technology*, New York, in press.

Polyakov, I., G.V. Alekseev, R.V. Bekryaev, U. Bhatt, R. Colony, M.A. Johnson, V.P. Karklin, D. Walsh, and A.V. Yulin (2003) Long-term ice variability in arctic marginal seas. *J. Climate, 16*(12), 2078-2085.

Rothrock, D.A., Y. Yu, and G.A. Maykut (1999) Thinning of the Arctic sea-ice cover. *GRL, 26*, 3469-3472.

Ocean

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Surface circulation regime

The circulation of the sea ice cover and ocean surface layer are closely coupled and are primarily wind-driven (Proshutinsky and Johnson, 1997). Data from satellites and drifting buoys indicate that the entire period of 2000-2006 has been characterized by an anticyclonic (clockwise) circulation regime due to a higher sea level atmospheric pressure over the region north of Alaska, relative to the 1948-2005 mean, and the prevalence of anticyclonic winds (Figure O1). Under these conditions, the clockwise circulation pattern in the Beaufort Sea region (the Beaufort Gyre) tends to be relatively strong. Conversely, in the cyclonic regime the clockwise circulation pattern in the Beaufort Sea region weakens, and the flow across the basin, from the Siberian and Russian coasts to Fram Strait (the Transpolar Drift), shifts poleward. The cyclonic pattern dominated during 1989-1996; the anticyclonic pattern has prevailed since 1997. The dominance of the anticyclonic regime during last decade of 1997-2006 is consistent with the Arctic Oscillation (AO) index (Figure A1) which fluctuated about zero indicating a relatively low level of influence from the Atlantic on these Arctic processes (Rigor et al., 2002).

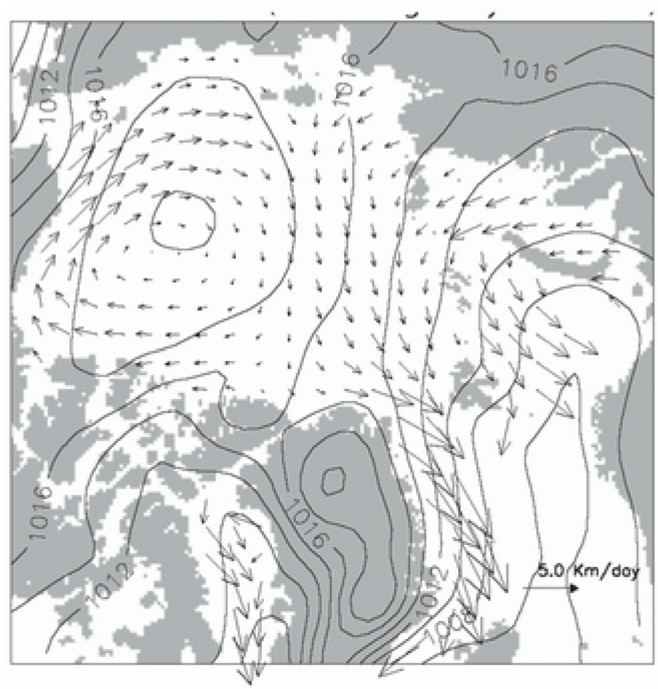

Figure O1. Sea ice drift pattern (arrows) in October-May 2000-2006 and sea surface atmospheric pressure distribution. Sea level atmospheric pressure is shown by lines (hPa) (courtesy of Ron Kwok).

Heat and freshwater content

From 2000 to 2006, intensive investigations have been conducted in the vicinity of the North Pole (North Pole Environmental Observatory, NPEO, http://psc.apl.washington.edu/northpole/) and in the Canada Basin (Beaufort Gyre Observing System, BGOS, http://www.whoi.edu/beaufortgyre/index.html). Observations show that in the previous decade (1990s) the water temperature and salinity fields of the Arctic Ocean changed dramatically relative to the climatology of the Environmental Working Group (EWG) Atlas of the Arctic Ocean (Arctic Climatology Project, 1997, 1998) where water temperature and salinity from observations were averaged and gridded for the decades of 1950, 1960, 1970 and 1980. Hydrographic data acquired at the North Pole in the 1990s show a strong increase in upper ocean salinity and a large increase in Atlantic Water temperature relative to EWG climatology.

From 2000 to 2005, the oceanographic conditions in the North Pole region relaxed to near the pre-1990 climatology (Figure O2). As characterized by average temperature and salinity anomalies relative to EWG climatology within 200 km of the North Pole, the change in the 1990s and the subsequent retreat to climatology are roughly consistent with a first order response to the AO with a 5-year time constant and 3-year time delay (Morison et al., 2006a). Recent results indicate conditions in 2006 at the Pole reverted to near 2004 conditions, but measurements of bottom pressure trends from 2002 to 2006 by the Gravity Recovery and Climate Experiment (GRACE) suggest a return of oceanographic conditions over the whole Arctic Ocean to pre-1990s conditions (Morison et al., 2006b). Preliminary 2007 data shows a slowing of this rate of return.

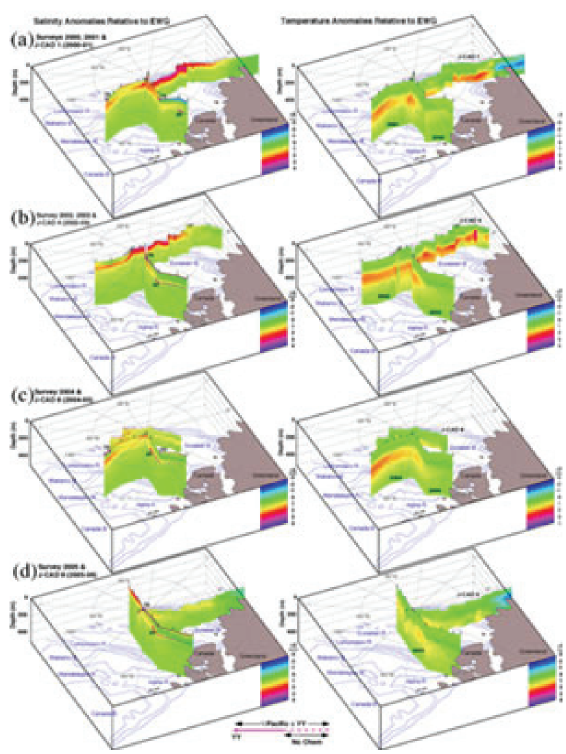

Figure O2. Salinity (l) and temperature (r) anomalies relative to EWG climatology along the NPEO surveys & JAMSTEC Compact Arctic Drifter (J-CAD) tracks for the years indicated on the temperature sections. Gray vertical lines mark survey station sites. Deep magenta lines (l) mark location of greater than 20% Pacific-derived water at 100-150 m. Surface lines mark greater than 70% Pacific-derived in the surface layer. From Morison et al., 2006a.

The Canada Basin hydrography in the 1990s has also changed relative to climatology but, in opposition to the salinity increase at the North Pole, the salinity of the upper layer in the Western Arctic was reduced. There are some indications that in the 2000s, relative to the 1990s, the salinity in this region has increased but it is still significantly less than in EWG climatology. Since 2000, the temperature of the Pacific and Atlantic waters in the Canada Basin is higher than in the 1990s and 0.8-1.0°C higher than in EWG climatology.

The Beaufort Gyre is the major reservoir of fresh water in the Arctic Ocean. In 2000-2006, the total freshwater content in the Beaufort Gyre has not changed dramatically relative to climatology but there is a significant change in the freshwater distribution (Figure O3, panels 3 and 4). The center of the freshwater maximum has shifted toward Canada and significantly intensified relative to climatology. Significant changes were observed in the heat content of the Beaufort Gyre (Figure O3, panels 1 and 2). It has increased relative to the climatology, primarily because of an approximately 2-fold increase of the Atlantic layer water temperature (Shimada et al., 2004). The Pacific water heat content in the Beaufort Gyre regions has also increased and it is possible that the pronounced sea ice reduction in this region, observed in 2006 (see Figure I1, right panel), resulted from heat released from this layer (Shimada et al., 2006). It is speculated that the major part of these changes in the freshwater and heat content occurred in the 1990s, but there are not enough data to confirm this.

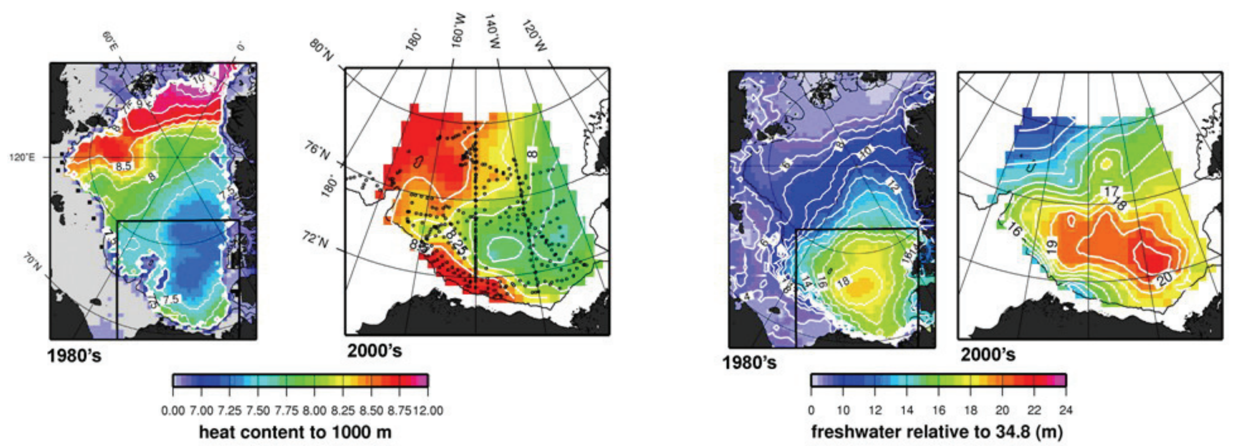

Figure O3. Summer heat (1.E^10 J/m^2, left) and freshwater (m, right) content. Panels 1 and 3 show heat and freshwater content in the Arctic Ocean based on 1980s climatology (Arctic Climatology Project, 1997, 1998). Panels 2 and 4 show heat and freshwater content in the Beaufort Gyre in 2000-2006 based on hydrographic surveys (black dots depict locations of hydrographic stations). For reference, this region is outlined in black in panels 1 and 3. The heat content is calculated relatively to water temperature freezing point in the upper 1000m ocean layer. The freshwater content is calculated relative to reference salinity of 34.8.

Sea Level

Figure O4 shows sea level time series from 9 coastal stations in the Siberian Seas (Arctic and Antarctic Research Institute data archives). These stations are still operational in the Arctic and have records for the period of 1954-2006. There is a positive sea level trend along arctic coastlines. Proshutinsky et al. (2004) estimated that for 1954-1989 the rate of sea level rise along arctic coastlines (40 stations), corrected for the glacial isostatic adjustment (GIA), was 0.185 cm/year. For the 9 stations shown in Figure O4 the rate for 1954-1989, after correction for their GIA, was 0.194 cm/year. Addition of 1990-2006 data increases the estimated rate of sea level change, beginning in 1954, to 0.250 cm/year.

The sea level time series correlates relatively well with the AO index and with the inverse of the sea level atmospheric pressure (SLP) at the North Pole. Consistent with these influences, sea level dropped significantly after 1990 and reached a minimum in 1996-1997 when the circulation regime changed from cyclonic to anticyclonic. In contrast, from 1997 to 2006 the mean sea level has generally increased in spite of the more or less stable behavior of AO and SLP. Since sea level change exhibits large interannual variability and is the net result of many individual effects of environmental forcing, it is difficult to evaluate the significance of the change in relative terms. The observed sea level rise during last 6-7 years could be related to decadal variability in combination with a general tendency of sea level to rise due to global warming (Greenland and Antarctic ice sheet melt) and, correspondingly, to the Arctic change expressed in an expansion of the water column due to increased water temperature (reduction of sea ice and solar warming in summer) and a decrease of water salinity (sea ice melt, increase of river runoff).

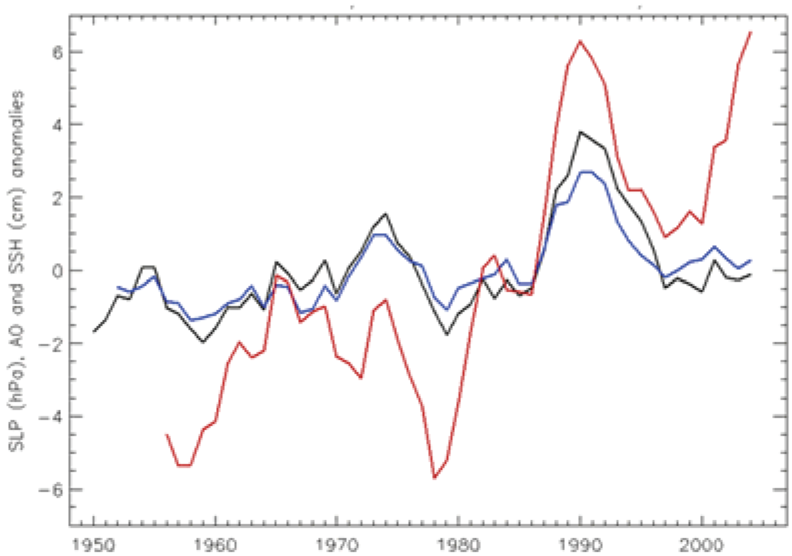

Figure O4. Annual mean anomalies of sea level at 9 tide gauge stations located along the Kara, Laptev, East-Siberian and Chukchi Sea coastlines (red). The blue line is the 5-year running mean anomalies of the annual mean Arctic Oscillation (AO) index multiplied by 3. The black line is the sea surface atmospheric pressure (SLP) anomaly at the North Pole (from from National Center for Atmospheric Research/NCEP reanalysis data) multiplied by -1.

References

Arctic Climatology Project (1997) Environmental Working Group joint U.S.-Russian atlas of the Arctic Ocean - winter period. Edited by L. Timokhov and F. Tanis. Ann Arbor, MI: Environmental Research Institute of Michigan in association with the National Snow and Ice Data Center. CD-ROM.

Arctic Climatology Project (1998) Environmental Working Group joint U.S.-Russian atlas of the Arctic Ocean - summer period. Edited by L. Timokhov and F. Tanis. Ann Arbor, MI: Environmental Research Institute of Michigan in association with the National Snow and Ice Data Center. CD-ROM.

Morison, J., M. Steele, T. Kikuchi, K. Falkner, and W. Smethie (2006a) Relaxation of central Arctic Ocean hydrography to pre-1990s climatology. *Geophys. Res. Lett., 33*, L17604, doi:10.1029/2006GL026826.

Morison, J., J. Wahr, R. Kwok, and C. Peralta-Ferriz, (2006b) Change in the Arctic Ocean as Observed with GRACE and In Situ Bottom Pressure Measurements. *Eos Trans. AGU, 87*(52), Fall Meet. Suppl., Abstract G12A-03 of presentation at the 2006 Fall AGU Meeting.

Proshutinsky, A., I.M. Ashik, E.N. Dvorkin, S. Häkkinen, R.A. Krishfield, and W.R. Peltier (2004) Secular sea level change in the Russian sector of the Arctic Ocean. *J. Geophys. Res., 109*, C03042, doi:10.1029/2003JC002007.

Proshutinsky, A.Y., and M.A. Johnson (1997) Two circulation regimes of the wind-driven Arctic Ocean. *J. Geophys. Res., 102*(C6), 12493-12514, 10.1029/97JC00738.

Rigor, I.G., J.M. Wallace, and R.L. Colony (2002) Response of Sea Ice to the Arctic Oscillation. *J. Climate, 15*, 2648-2663.

Shimada K., T. Kamoshida, M. Itoh, S. Nishino, E. Carmack, F.A. McLaughlin, S. Zimmermann, and A. Proshutinsky (2006) Pacific Ocean inflow: Influence on catastrophic reduction of sea ice cover in the Arctic Ocean. *Geophys. Res. Lett., 33*, L08605, doi:10.1029/2005GL025624.

Shimada, K., F. McLaughlin, E. Carmack, A. Proshutinsky, S. Nishino, and M. Itoh (2004) Penetration of the 1990s warm temperature anomaly of Atlantic Water in the Canada Basin. *Geophys. Res. Lett., 31*, L20301, doi:10.1029/2004GL020860.

Land

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Permafrost

Long-term permafrost temperature data are available only from a few clusters of stations, mostly in North America. Observations from the long-term sites show a general increase in permafrost temperatures during the last several decades in Alaska (Osterkamp and Romanovsky, 1999; Romanovsky et al., 2002; Osterkamp, 2003), northwest Canada (Couture et al., 2003; Smith et al., 2003), Siberia (Pavlov, 1994; Oberman and Mazhitova, 2001; Romanovsky et al., 2002; Pavlov and Moskalenko, 2002), and Northern Europe (Isaksen et al., 2000; Harris and Haeberli, 2003). Uninterrupted permafrost temperature records for more than a 20-year period have been obtained by the University of Alaska Fairbanks along the International Geosphere-Biosphere Programme Alaskan transect, which spans the entire continuous permafrost zone in the Alaskan Arctic. All of the observatories show a substantial warming during the last 20 years. This warming was different at different locations, but was typically from 0.5 to 2°C at the depth of zero seasonal temperature variations in permafrost (Osterkamp, 2005). In 2006, there was practically no change to the mean annual temperatures at the permafrost surface if compared to 2005 (Romanovsky et al., 2006). These data also indicate that the increase in permafrost temperatures is not monotonic. During the observational period, relative cooling has occurred in the mid-1980s, in the early 1990s, and then again in the early 2000s. As a result, permafrost temperatures at 20 m depth experienced stabilization and even a slight cooling during these periods.

Very similar permafrost temperature dynamics were observed in the European North of Russia during the same period (Figure L1). However, there is some lag in the soil temperature variations at the Alaskan sites compared to the Russian sites. This observation is similar to what was discovered in comparison of permafrost temperature dynamics in Fairbanks, Alaska and Yakutsk, Russia (Romanovsky et al., 2007). Relative cooling has occurred in Vorkuta region in the early and late 1980s and then in late 1990s. The total warming since 1980 was almost 2°C at the Vorkuta site.

Data on changes in the active layer thickness (ALT) in the arctic lowlands are less conclusive. In the North American Arctic, ALT experiences a large interannual variability, with no discernible trends. This is likely due to the short length of historical data records (Brown et al., 2000). A noticeable increase in the active layer thickness was reported for Mackenzie Valley (Nixon et al., 2003). However, this positive trend was reversed into a negative trend at the most of these sites after 1998 (Tarnocai et al., 2004). An increase in thickness of more than 20 cm between the mid-1950s and 1990 derived from the historical data collected at the Russian meteorological stations was reported for the continuous permafrost regions of the Russian arctic (Frauenfeld et al., 2004; Zhang et al., 2005). At the same time, reports from several specialized permafrost research sites in Central Yakutia show no significant changes in the active layer thickness (Varlamov et al., 2001; Varlamov, 2003). The active layer was especially deep in 2005 in Interior

Alaska. Around Fairbanks the 2005 active layer depth was the deepest observed in the past 10 years. Data from many of these sites show that the active layer developed during the summer of 2004 (one of the warmest summers in Fairbanks on record) did not completely freeze during the 2004-2005 winter. A thin layer just above the permafrost table was unfrozen during the entire winter. Active layer in the summer of 2006 was also one of the deepest on record at most of our observation sites in the Fairbanks area even though the summer air temperatures were close to normal.

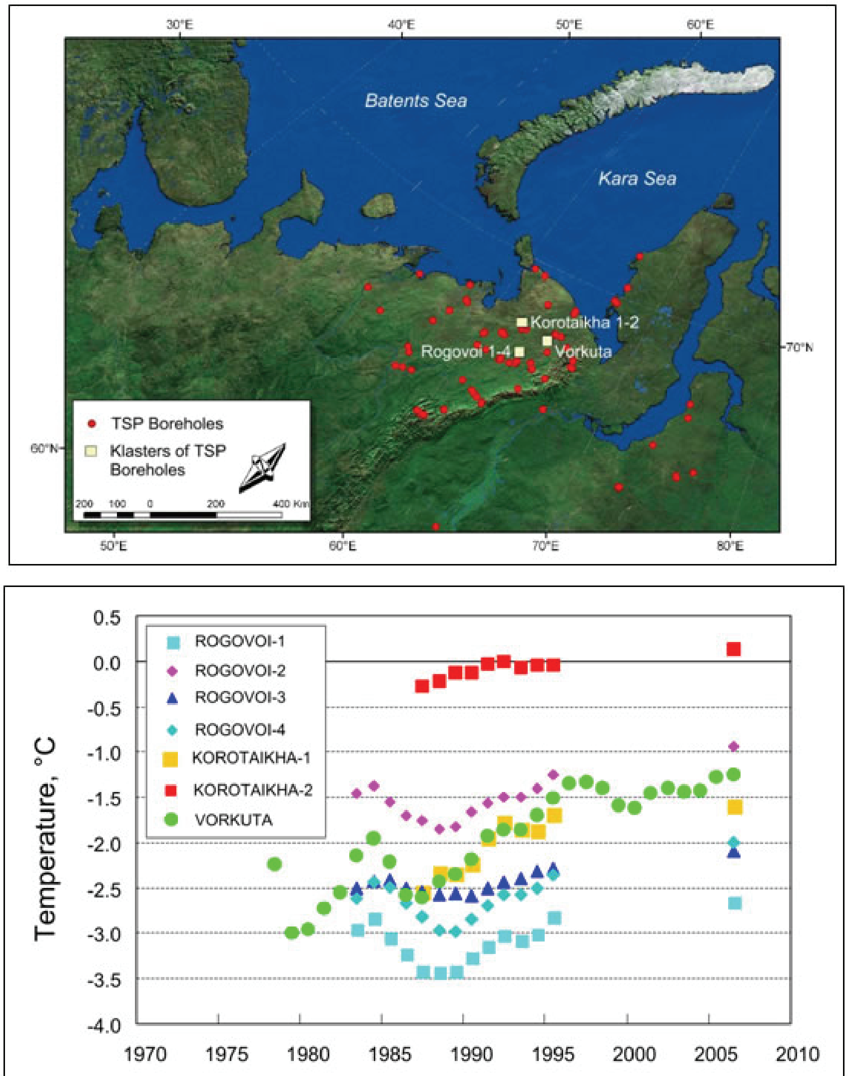

Figure L1. Top: Location of the long-term MIREKO permafrost observatories in northern Russia. Bottom: Changes in permafrost temperatures at 15 m depth during the last 20 to 25 years (Oberman, 2007).

Thermokarst (Thermokarst is a land surface formed as permafrost melts)

Thermokarst topography forms as ice-rich permafrost thaws, either naturally or anthropogenically, and the ground surface subsides into the resulting voids. The important and dynamic processes involved in thermokarsting include thaw, ponding, surface and subsurface drainage, surface subsidence and related erosion. These processes are capable of rapid and extensive modification of the landscape. Recent analysis suggests that in regions over thin permafrost (~<20 m), surface ponds may shrink and surface soils may become drier as the permafrost degrades (Figure L2). In colder regions with thicker permafrost, as the warming proceeds, near surface ice thaws, the land surface subsides and new water bodies are formed (Hinkel et al., 2007; Jorgenson and Shur, 2007; Jorgenson et al., 2006; Riordan et al., 2006; Smith, et al., 2005; Yoshikawa and Hinzman, 2003).

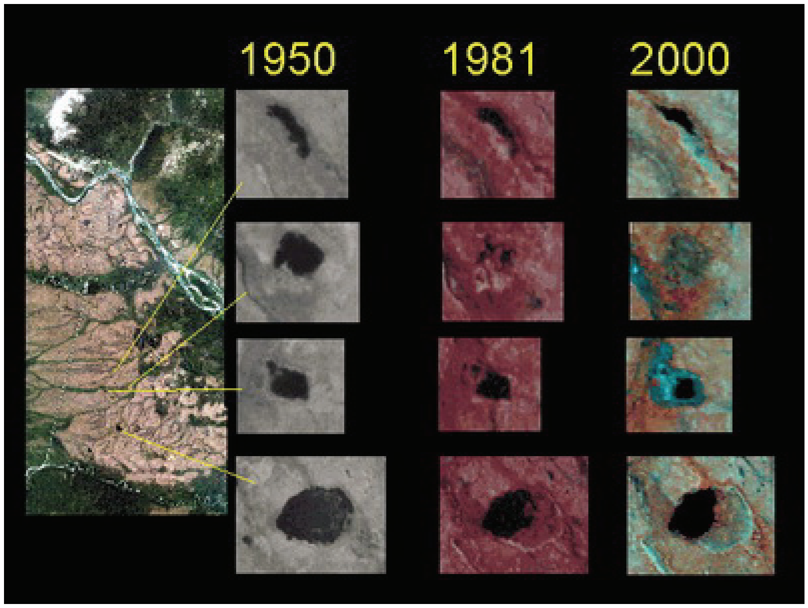

Figure L2. Numerous tundra ponds near Council, Alaska (64° 51' N, 163° 42' W) have decreased in surface area over the last 50 years. A probable mechanism for these shrinking ponds is internal drainage through the degradation of shallow permafrost (Yoshikawa and Hinzman, 2003).

Extensive thermokarsting, resulting in the creation of new water-filled surface depressions, has recently been observed on the Beaufort Coastal Plain in northern Alaska (Jorgenson et al. 2006). Analysis of aerial photography indicated that widespread ice wedge degradation had not occurred before 1980. Field observations and sampling showed that ice wedge degradation has been relatively recent, as indicated by newly drowned vegetation. Despite the relatively cold average annual temperature of this northern permafrost, thermokarst was widespread on a variety of terrain conditions, but most prevalent on ice-rich centers of old drained lake basins and alluvial-marine terraces.

The ponds on the Seward Peninsula were examined to determine if recent changes in climate have impacted the dynamics of their development and degradation (Yoshikawa and Hinzman, 2003). Of the 24 ponds studied, 22 have decreased in area between 1951 and 2000.

Smith et al., 2006 demonstrated that similar processes are occurring in the Siberian Arctic. In northern Siberia, where cold continuous permafrost dominates, there has been a significant increase in the number and size of surface water bodies (consistent with the Jorgenson et al, 2006 study). In the more southerly parts of Siberia, where warmer, discontinuous permafrost predominates, there has been a significant decrease in the number and size of surface water bodies (consistent with the Yoshikawa and Hinzman, 2003 study).

Snow Extent

For the Northern Hemisphere winter of 2005-2006, the microwave data indicate negative departures from the long term mean (1978-2006) for every month except March with an average negative departure for the winter months (November through April) of approximately 1.3 million km^2. For the calendar year of 2006, the NOAA data indicate an average snow cover extent of 24.9 million km^2 which is 0.6 million km^2 less than the 37-year average (Robinson, personal communication).

Northern Hemisphere snow cover extent has a mean maximum of approximately 47 million km^2, typically in February. The minimum usually occurs in August and is less than about 1 million km^2, most of which is snow on glaciers and perennial snow fields. As a result, snow cover is the land surface characteristic responsible for the largest annual and interannual differences in land surface albedo (Figure L3). Snow covers a much smaller area in the Southern Hemisphere and plays a relatively small role in global climate.

Hemispheric-scale snow cover fluctuations are monitored with satellite data. Since 1966, NOAA has produced snow extent charts (Robinson et al., 1993; Frei and Robinson, 1999). These charts were primarily derived from the manual interpretation of visible band imagery until 1999, when passive microwave and other data sources became available (Ramsay, 1998; NOAA/NESDIS/OSDPD/SSD, 2004, update 2006). Passive microwave data can enhance snow measurements based on visible data alone, sensing the surface through clouds and in darkness. However, passive microwave may not detect some areas of shallow snow that can be seen in visible band imagery. As a result, time series from the two sources can differ. Figure L4 compares a microwave data derived snow cover data set (Armstrong and Brodzik, 2001; Armstrong et al., 2005b) with NOAA snow extent data. Both show similar inter-annual variability and consistently indicate Northern Hemisphere maximum extents exceeding 40 million km^2. The NOAA time series indicates a decreasing trend of -2.0% per decade (Brodzik et al. 2006). There is a decreasing trend of -0.7% per decade in the microwave snow cover, although it is not significant at the 90% level. Both sources indicate a decreasing trend in snow cover in every month but November and December. The strongest seasonal signal occurs during May to August when both indicate significant decreasing trends. The western United States is among the regions with the strongest decreasing trends, supporting Groisman et al. (2004) and Mote et al. (2005) results using in situ observations. Shallow snow cover at low elevations in temperate regions is the most sensitive to temperature fluctuations and hence most likely to decline with increasing temperatures (IPCC 2007).

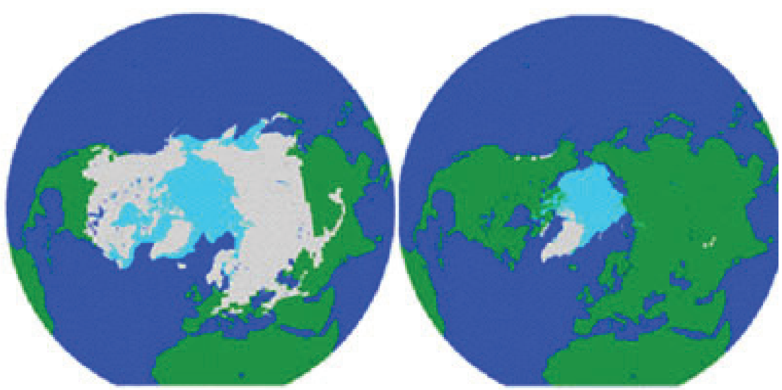

Figure L3. Mean snow cover extent (grey), 1966-2006, for February (left) and August (right) from the Northern Hemisphere Equal Area Scalable Earth (EASE)-Grid Weekly Snow Cover and Sea Ice Extent data set (Armstrong and Brodzik, 2005). The product includes climatologies of snow average conditions, probability of occurrence, and variance based on NOAA charts as revised by Robinson et al. (1993).

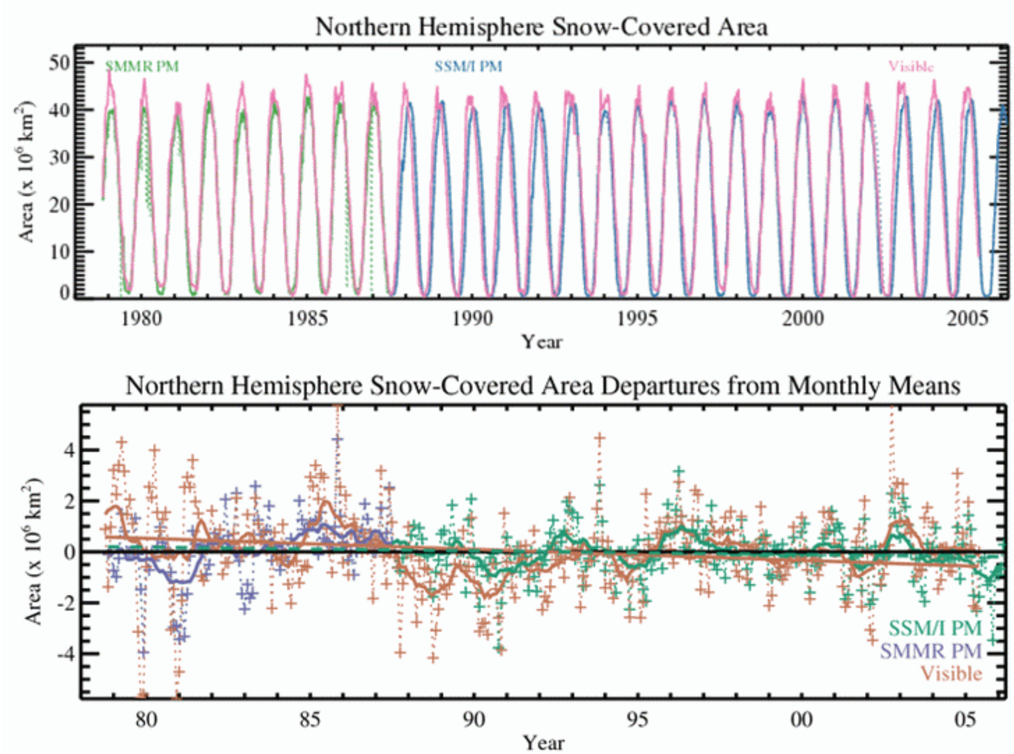

Figure L4. Time series of Northern Hemisphere snow-covered area (SCA) derived from passive microwave (green/blue) and visible (pink) sensors (top), and SCA departures from monthly means (bottom), from NOAA snow charts (orange) and microwave (purple/green) data sets.

Glaciers

Glaciers and ice caps, excluding those adjacent to the large ice sheets of Greenland and Antarctica, can be found on all continents except Australia and have an estimated area between 512 and 540 x 10^3 km^2. The complicated and uncertain processes that control how fast glaciers move make it difficult to use changes in the areal extent of glaciers as a straightforward indicator of changes in climatic conditions. Further, many large collections of glacier photographs are available, but it is only in the last decade or so that remote sensing imagery has provided a means to monitor changes in the areal extent of glaciers. The Global Land Ice Measurements from Space glacier database project, with participation from more than 60 institutions in 28 nations, is working now on a baseline study to quantify the areal extent of existing glaciers (Armstrong et al., 2005a).

Mass balance measurements, or the difference between the accumulation and ablation, are a more direct method to determine the year-to-year "health" of a glacier. Changes in mass balance correspond to changes in glacier volume. These measurements are typically obtained from less than about 0.2 percent of the world's glaciers. Researchers have measured mass balance on more than 300 glaciers since 1946, although a continuous record exists for only about 40 glaciers since the early 1960s. Nevertheless, considerable compilation and analysis has occurred (e.g. Cogley, 2005). These results indicate that in most regions of the world, glaciers are shrinking in mass. From 1961 to 2003, the thickness of "small" glaciers decreased approximately 8 meters, or the equivalent of more than 6,000 cubic kilometers of water (see http://nsidc.org/sotc/glacier_balance.html). Recent mass loss of glaciers and ice caps is estimated to be 0.51+/- 0.32 mm sea level equivalent (SLE) per year between 1961 and 2003 and 0.81 +/-0.43 mm SLE per year between 1993 and 2003 (Dyurgerov and Meier, 2006; http://nsidc.org/sotc/sea_level.html). The greatest mass losses per unit area are found in Patagonia, Alaska and NW USA/SW Canada. However, because of the corresponding large areas, the biggest contributions in total to sea level rise come from Alaska, the Arctic and the Asian high mountains.

River discharge

The river discharge database R-ArcticNet (http://www.R-Arcticnet.sr.unh.edu) was extended up to 2004 for 48 downstream river gauges. The last five years were characterized by an increase of total discharge to the Arctic Ocean mainly due to a contribution from Asian rivers. Mean 2000-2004 discharge from Asia was 110 km^3 (5%) higher than over the previous twenty years. The mean discharge to the ocean from North America and Europe for 2000-2004 was practically unchanged relative to 1980-1999. A consistent increase in river discharge is observed from Eurasia for a longer time interval as well. Mean discharge over 2000-2004 for the large Eurasian rivers was 3-9% higher than the discharge over 1936-2004. Thus the contemporary data further confirms the presence of a significant increasing trend in the fresh water discharge to the Arctic Ocean from Eurasia documented earlier by Peterson et al. (2002). The maximum total discharge of the six largest Eurasian rivers over 1936-2004 was observed in 2002, at 2080 km^3/year (see BAMS State of the Climate Report, 2006).

References

Armstrong, R.L., and M.J. Brodzik (2005) Northern Hemisphere EASE-Grid weekly snow cover and sea ice extent version 3, Boulder, Colorado, National Snow and Ice Data Center. Digital media.

Armstrong, R.L., B. Raup, S.J.S. Khalsa, R. Barry, J. Kargel, C. Helm, and H. Kiefer (2005a) GLIMS glacier database. National Snow and Ice Data Center, Boulder, Colorado, digital media. [Available online at http://nsidc.org/data/nsidc-0272.html]

Armstrong, R.L, M.J. Brodzik, K. Knowles, and M. Savoie (2005b) Global monthly EASE-Grid snow water equivalent climatology. National Snow and Ice Data Center, Boulder, Colorado, digital media. [Available online at http://nsidc.org/data/nsidc-0271.html]

Brodzik, M.J., R.L. Armstrong, E.C. Weatherhead, M.H. Savoie, K.W. Knowles, and D.A. Robinson (2006) Regional trend analysis of satellite-derived snow extent and global temperature anomalies. *Eos Trans. AGU, 87*(52), Abstract U33A-0011.

Brown, J., K.M. Hinkel, and F.E. Nelson (2000) The Circumpolar Active Layer Monitoring (CALM) program: Research Designs and Initial Results. *Polar Geogr., 24*,163-258.

Cogley, G.J. (2005) Mass and energy balances of glaciers and ice sheets. In: M. Anderson (Editor), *Encyclopedia of Hydrological Sciences*, John Wiley & Sons, Ltd.

Couture R., S. Smith, S.D. Robinson, M.M. Burgess, and S. Solomon (2003) On the hazards to infrastructure in the Canadian North associated with thawing of permafrost. Proceedings of Geohazards 2003, 3rd Canadian Conference on Geotechnique and Natural Hazards. The Canadian Geotechnical Society: Edmonton, Alberta, Canada; 97–104.

Dyurgerov, M., and M.F. Meier (2006) Glaciers and changing earth system: a 2004 snapshot, *Data of glaciological studies (MGI), 100*, (in press).

Frauenfeld, O.W., T. Zhang, R.G. Barry, and D. Gilichinsky (2004) Interdecadal changes in seasonal freeze and thaw depths in Russia. *J. Geophys. Res., 109*, D05101, doi:10.1029/2003JD004245.

Frei, A., and D.A. Robinson (1999) Northern Hemisphere snow extent: regional variability 1972-1994. *Int. J. Climatology, 19*, 1535-1560.

Groisman, P. Ya., T.R. Karl, D.R. Easterling, B. Sun, and J.H. Lawrimore (2004) Contemporary changes of the hydrological cycle over the contiguous United States: trends derived from in situ observations. *J. Hydrometeorol., 5*, 64-85.

Harris, C., and W. Haeberli (2003) Warming permafrost in European mountains. *World Meteorol. Org. Bull., 52*(3), 6 pp., see also Global and Planetary Change, 39(2003), 215-225.

Hinkel, K.M., B.M. Jones, W.R. Eisner, C.J. Cuomo, R.A. Beck, and R. Frohn (2007). Methods to assess natural and anthropogenic thaw lake drainage on the western Arctic coastal plain of northern Alaska, *J. Geophys. Res., 112*, F02S16, doi:10.1029/2006JF000584.

IPCC (2007) Working Group 1 Fourth Assessment Report: The Physical Science Basis of Climate Change, Chapter 4: Observations: Changes in Snow, Ice and Frozen Ground.

Isaksen, K., D. Vonder Mühll, H. Gubler, T. Kohl, and J.L. Sollid (2000) Ground surface temperature reconstruction based on data from a deep borehole in permafrost at Janssonhaugen, Svalbard. *Ann. Glaciol., 31*, 287-294.

Jorgenson, M.T., and Y. Shur (2007). Evolution of lakes and basins in northern Alaska and discussion of the thaw lake cycle, *J. Geophys. Res., 112*, F02S17, doi:10.1029/2006JF000531.

Jorgenson, M.T., Y.L. Shur, and E R. Pullman (2006). Abrupt increase in permafrost degradation in Arctic Alaska, *Geophys. Res. Lett., 33*, L02503, doi:10.1029/2005GL024960.

Mote, P.W., A.F Hamlet, M.P. Clark, and D.P. Lettenmaier (2005) Declining mountain snowpack in western North America. *Bull. of Amer. Meteorol. Soc., 86*, 39-49, doi:10.1175/BAMS-86-1-39.

Nixon, F.M., C. Tarnocai, and L. Kutny (2003) Long-term active layer monitoring: Mackenzie Valley, northwest Canada, in Permafrost, Phillips, Springman & Arenson, pp. 821-826.

NOAA/NESDIS/OSDPD/SSD (2004, update 2006) *IMS daily Northern Hemisphere snow and ice analysis at 4 km and 24 km resolution*. Boulder, CO: National Snow and Ice Data Center. Digital media.

Oberman, N.G. (2007) Some peculiarities in permafrost degradation in the Pechora-Urals region in Russia, In Proceedings of the International Conference: *Cryogenic Resources of Polar Regions*, Syktivkar, Russia, June 17-20, 2007, in press.

Oberman, N.G. and G.G. Mazhitova (2001) Permafrost dynamics in the northeast of European Russia at the end of the 20th century. *Norw. J. Geog., 55*, 241-244.

Osterkamp, T.E. (2005) The recent warming of permafrost in Alaska, *Global Planet. Change, 49*: 187-202.

Osterkamp, T.E. (2003) A thermal history of permafrost in Alaska, Proceedings of the 8th Internaional Conference on Permafrost, July 21-25, 2003, Zurich, Switzerland, Phillips, M., S.M. Springman, and L. U. Arenson (eds.), A. A. Balkema, Lisse, The Netherlands, Vol. 2, pp. 863-868.

Osterkamp, T.E. (2005) The recent warming of permafrost in Alaska. *Global Planet. Change, 49*, 187-202.

Osterkamp, T.E., and V.E. Romanovsky (1999) Evidence for warming and thawing of discontinuous permafrost in Alaska. *Permafrost Periglac., 10*(1), 17-37.

Pavlov, A.V. (1994) Current changes of climate and permafrost in the Arctic and Sub-Arctic of Russia. *Permafrost Periglac., 5*, 101-110.

Pavlov, A.V., and N.G. Moskalenko (2002) The thermal regime of soils in the north of Western Siberia. *Permafrost Periglac., 13*(1), pp. 43-51.

Peterson, B.J., R.M. Holmes, J.W. McClelland, C.J. Vörösmarty, R.B. Lammers, A.I. Shiklomanov, I.A. Shiklomanov, and S. Rahmstorf (2002) Increasing river discharge to the Arctic Ocean. *Science, 298*, 2171-2173.

Ramsay, B.H. (1998) The Interactive Multisensor Snow and Ice Mapping System. *Hydrol. Processes, 12*, 1537-1546.

Riordan, B., D. Verbyla, and A.D. McGuire (2006). Shrinking ponds in subarctic Alaska based on 1950-2002 remotely sensed images, *J. Geophys. Res., 111*, G04002, doi:10.1029/2005JG000150.

Robinson, D.A., K.F. Dewey, and R.R Heim (1993) Global snow cover monitoring: an update. *Bull. Am. Meteorol. Soc., 74*, 1689-1696.

Romanovsky, V.E., M. Burgess, S. Smith, K. Yoshikawa, and J. Brown (2002) Permafrost temperature records: Indicator of climate change. *EOS, 83*(50), 589, 593-594.

Romanovsky, V.E., S.S. Marchenko, R. Daanen, D. Nikolsky, D.O. Sergeev, and D.A. Walker (2006) Soil climate and frost heave along the Permafrost/Ecological North American Arctic Transect. *Eos Trans. AGU, 87*(52), Fall Meet. Suppl., C44A-06.

Romanovsky, V.E., Sazonova, T.S., Balobaev, V.T., Shender, N. I., and D.O. Sergueev (2007) Past and recent changes in permafrost and air temperatures in Eastern Siberia. *Environmental and Global Change*, in press.

Smith, L.C., Y. Sheng, G.M. MacDonald, and L.D. Hinzman (2005). Disappearing Arctic Lakes, *Science, 308*:1429, 3 June 2005.

Smith, S. L., Burgess, M. M., and A.E. Taylor (2003) High Arctic permafrost observatory at Alert, Nunavut – analysis of e 23 year data set. In: Proceedings of the 8th Internaional Conference on Permafrost, July 21-25, 2003, Zurich, Switzerland, Phillips, M., S. M. Springman, and L. U. Arenson (eds.), A. A. Balkema, Lisse, The Netherlands, Vol. 2, pp. 1073-1078.

Tarnocai, C., F.M. Nixon, and L. Kutny (2004) Circumpolar-Active-Layer-Monitoring (CALM) Sites in the Mackenzie Valley, Northwestern Canada, *Permafrost Periglac., 15*, 141-153.

Varlamov, S.P. (2003) Variations in the thermal state of the lithogenic base of landscapes in Central Yakutia, Proceedings of the Second International Conference "The Role of Permafrost Ecosystems in Global Climate Change", 12-17 August 2002, Yakutsk, Russia, pp. 52-56.

Varlamov, S.P., Skachkov, Yu. B., Skryabin, P.N. and N.I. Shender (2001) Thermal response of the lithogenic base of permafrost landscapes to recent climate change in Central Yakutia, Proceedings of the International Conference "The Role of Permafrost Ecosystems in Global Climate Change", 3-5 May 2000, Yakutsk, Russia, pp. 44-45.

Yoshikawa, K. and L.D. Hinzman (2003). Shrinking thermokarst ponds and groundwater dynamics in discontinuous permafrost, *Permafrost Periglac., 14*(2):151-160.

Zhang, T., O. W. Frauenfeld, M.C. Serreze, A.J. Etringer, C. Oelke, J.L. McCreight, R.G. Barry, D. Gilichinsky, D. Yang, H. Ye, F. Ling, and S. Chudinova (2005) Spatial and temporal variability of active layer thickness over the Russian Arctic drainage basin. *J. Geophys. Res., 110*, D16101, doi:10.1029/2004JD005642.

Greenland Ice Sheet Mass Balance

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Summary

Recent relatively high summer temperatures (1995-2005) are associated with increased net ice loss over Greenland. Recent warm events are about the same magnitude, if not smaller, than those of the early 20th century warm period (1918-1947). 2006 was not as warm as other recent years such as 2003 or 2005. Physical response mechanisms, such as hydraulic acceleration of the ice sheet from continued warming, remain incompletely understood.

Satellite Observations

The Greenland ice sheet (GrIS) contains 7.4m global sea-level equivalent and is vulnerable to mass loss by climate warming (Gregory et al. 2004). Observational studies have provided many insights into recent GrIS mass balance changes. Airborne and satellite laser-altimetry data analyses indicate a volume loss of about 60 $km^3\ yr^{-1}$ in the 1993/4 - 1998/9 period, that increased to about 80 $km^3\ yr^{-1}$ in 1997-2003 (Krabill et al. 2000, 2004, Thomas et al. 2006). Various recent analyses of gravimetric (GRACE) satellite data suggest greater mass (volume) losses in the 101-226 Gt yr^{-1} (111-248 $km^3\ yr^{-1}$) range within the recent few years, that is, 2002-2006 (Chen et al. 2006, Luthcke et al. 2006, Ramillien et al. 2006, Velicogna and Wahr 2006). The largest mass losses are generally indicated as being from low elevations (<~2000 m) and especially in southeast Greenland, with partly compensating mass gains at higher elevations (>2000 m) (Luthcke et al. 2006). Other altimetry data suggests that there appears to have been substantial growth of the GrIS interior from 1992-2003 (Johannessen et al. 2005, Thomas et al. 2005, Zwally et al. 2005), which may be partly attributable to increased atmospheric moisture and precipitation and/or shifting storm tracks forced by enhanced greenhouse gases (IPCC 2007).

Satellite radar interferometry (InSAR) reveals widespread acceleration of Greenland glaciers, a pattern progressing northward since 1996, with an accompanying doubling of the ice sheet's volume deficit from approximately 90 to 220 $km^3\ yr^{-1}$ (Rignot and Kanagaratnam 2006), although it is as yet too early to say just how exceptional are these changes. The recent accelerations of Kangerdlussuaq and Helheim (outlet) glaciers appear to be driving the recent (2001-2006) concentration of mass loss in south-east Greenland (Stearns and Hamilton 2007); however, the velocities of these glaciers appear to have decreased again in 2006 to near their previous rates (Howat et al. 2007). Some of these margin changes may signal a response to recent climatic warming through infiltration of surface meltwater in a basal dynamic positive feedback mechanism (Zwally et al. 2002, Parizek and Alley 2004). Alternatively, thinning and breakup of Jakobshavn Glacier's (SW Greenland) floating ice tongue and acceleration of the glacier itself—as well as that of Helheim Glacier in south-east Greenland—may have been induced, or at least influenced, by forcing from the ocean (Thomas et al. 2003, Howat et al. 2005, Bindschadler 2006).

Taking all approaches together, the general consensus is an accelerating mass loss of the GrIS over the past decade 1995-2005 (IPCC 2007 Ch. 4). A recent survey concludes that the GrIS is currently losing ~100 Gt yr^{-1} (Shepherd and Wingham 2007). However, there remains considerable discrepancy among these pioneering observational estimates. Most of the observational studies have data spans of less than a decade, which also means that the interpretation of their results may be seriously affected by large year-to-year variability in GrIS mass turnover, e.g. sudden glacier accelerations (Rignot and Kanagartnam 2006). Since the Helheim and Kangerdlugsuaq glaciers have been shown to accelerate and decelerate over just a few years (Howat et al. 2007), these 'speed-ups' may just represent flow variability on interannual time scales and therefore represent the 'weather' rather than the 'climate' of the GrIS (Shepherd and Wingham 2007, Howat et al. 2007). All in all, we cannot be certain that the apparently accelerating GrIS mass loss and unexpected, rapid changes in its outlet glaciers, represent a profound change in ice-sheet behaviour (Murray 2006). The above studies alone, therefore, have yet to provide a convincing broad temporal perspective on how the GrIS might be responding to long-term climatic change, most notably the evident global and regional warming since the 1970s (IPCC 2007).

Surface Mass Balance

Surface mass balance (SMB), essentially net snow accumulation minus meltwater runoff, time series are available (e.g., Hanna et al. 2005; Box et al. 2006), which can help place the remotely sensed results into a longer, multi-decadal, climatic perspective. These series are based on meteorological models that assimilate observations for calibration and verification. There is good agreement of respective annual precipitation and runoff values from the two independently-derived SMB series for the period of published overlap (1988-2004). However, by definition, neither SMB series takes into account the mass losses from iceberg calving and basal melting, for which only crude estimates can currently be given. Iceberg calving is roughly equivalent to the amount of annual runoff whereas the basal melting is probably relatively small for Greenland.

Polar atmospheric model calculations after Box et al. (2006) suggest that 2006 was a normal year for snow accumulation for the ice sheet as a whole (Table 1), accumulation being the difference of solid plus liquid precipitation and surface-water-vapor exchanges. A positive 2006 meltwater production anomaly, however, exceeded one standard deviation of the 1988-2004 mean. The extreme year 2005 was excluded to facilitate comparison with Box et al. (2006) and does not significantly change our conclusion that 2006 had likely abnormally negative SMB. Runoff rates were apparently 57% above this recent 17-year mean. The accumulation-area ratio was below the value of recent decades, owing to slightly below normal precipitation and an apparent expansion of the ablation zone. Enhanced 2006 melt rates are also suggested by above normal temperatures, in particular for the southern part of the ice sheet south of 64° N, where MODIS satellite data processed using the Liang et al. (2005) method suggest low-albedo (dark surface) anomalies relative to the 2000-2006 base period.

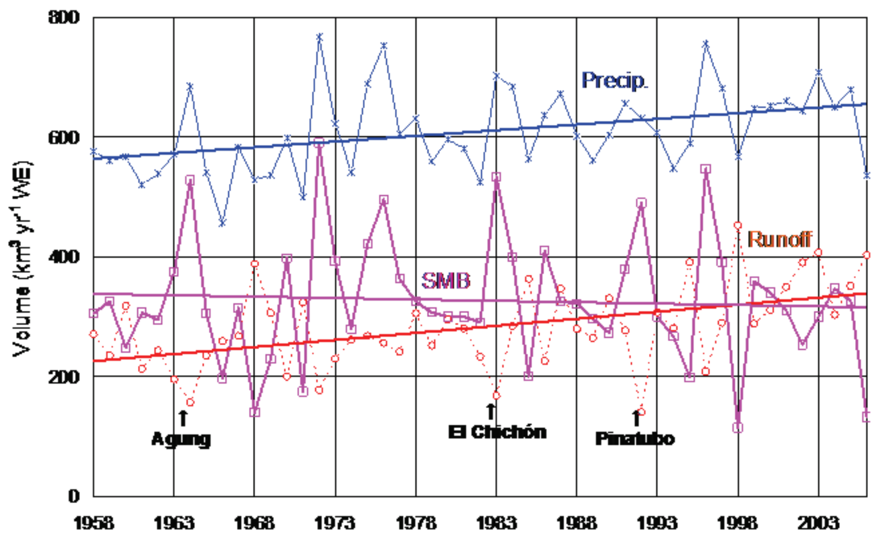
Table 1. Polar MM5 Greenland ice sheet surface mass balance parameters: 2006 departures from 1988-2004 average

	% of 1988-2004 average	2006 minus 1988-2004 average [km^3 y^{-1}]*
Total precipitation	-2%	-11
Liquid precipitation	-4%	-1
Evaporation	0%	0
Blowing snow sublimation	0%	0
Snow accumulation	-2%	-12
Meltwater production	38%	71
Meltwater runoff	57%	49
Surface mass balance	N/A	-60
Mean temperature	N/A	2.7K
Accumulation area ratio	-3.8%	-0.035 (ratio)

* unless otherwise indicated

A 49-year (1958-2006) SMB series, updated and recalibrated from Hanna et al. (2005), reveals 1998, 2003 and 2006 as, respectively, the first, second and third highest runoff years (Fig. 1). There is a statistically significant underlying trend-line increase in runoff from 1958-2006 of 113.0 km^3 yr^{-1} (equivalent to 40.0% of mean 1958-2006 runoff, compared with a standard deviation of 24.3% of the mean). Moreover, the five highest runoff years have all occurred since and including 1995, and five of the nine highest runoff years since 2001 inclusive. Greenland ice sheet precipitation follows a significantly increasing trend of 90.9 km^3 or +14.9% over this 49 year period, compared with a standard deviation of 69.7 km^3 yr^{-1} or 11.4% (Fig. 1). Additional precipitation, mainly in the form of snow accumulation, therefore largely (about three quarters) offsets rising Greenland runoff in terms of the SMB. There is thus a relatively small and insignificant negative trend in SMB of -22.1 km^3 yr^{-1} (sigma = 104.8 km^3 yr^{-1}) from 1958-2006, highlighting the sensitive balance between increased snow accumulation in the interior of the ice sheet and increased runoff around the edges. The further mass loss from ice dynamics due to accelerated flow of outlet glaciers was at least several times larger for the recent few warmest years (Rignot and Kanagaratnam 2006).

Figure 1. Greenland ice sheet precipitation, surface meltwater runoff and surface mass balance (SMB = solid precipitation minus evaporation minus runoff) series for 1958-2006, recalibrated and updated from Hanna et al. (2005). Note significantly increasing precipitation and runoff trends but negligible SMB change. We may conclude from the trend analysis that the hydrological system of the ice sheet has become more vigorous, i.e. with more mass turnover at the surface. The dynamic response to the increased turnover remains critical to understand.



Recent high snow accumulation events occurred in winter 2004/05, concentrated in west Greenland (Nghiem et al. 2007), and winter-spring 2002/03 in SE Greenland (Krabill et al. 2004; Box et al. 2005; Hanna et al. 2006). On the other hand, 2006 was the sixth lowest precipitation year in the 49-year ECMWF Greenland record after Hanna et al. (2006), which together with the high 2006 runoff, resulted in the second-lowest annual ice sheet net mass input (SMB) since 1958.

Notably, our SMB data do not agree with model predictions that suggest increased Greenland accumulation may be outweighed by rising runoff in a warmer climate (Huybrechts et al. 2004). However, according to the mechanism proposed by Zwally et al. (2002), the additional meltwater that we observe may already be more readily reaching the bed of the GrIS and prompting accelerated flow of Greenland outlet glaciers—likely a substantial fraction of the enhanced flow detected by Rignot and Kanagaratnam (2006). Such amplification might explain the significant increases in overall mass loss strongly suggested by the consensus of GrIS mass balance estimates for the past decade (IPCC 2007 Ch. 4) but is far from proven. However, since some recent key glacier accelerations in east Greenland have already subsided (Howat et al. 2007), the sustainability of the enhanced flow is questionable. If the Zwally effect is dominant (which very much remains to be shown), then flow variability is of course directly connected to surface-runoff variability.

Surface Air Temperature

Southern Greenland temperature changes since the early 1990s in summer reflect general northern hemisphere and global warming. This differs from the Greenland warming phase between 1918 and 1947 when there was less apparent linkage between Greenland and the northern hemisphere average. According to a composite record of seven coastal Greenland stations south of 70° N latitude, summer 2003 was the warmest since 1958 (1958 marks the start of the European Centre for Medium-Range Weather Forecasts (ECMWF) ERA-40 climate reanalysis and hence the Hanna et al. (2005, 2007) modelled surface mass balance series). The second warmest summer, in 2005, had the most extensive anomalously warm conditions over the ablation zone of the ice sheet. Spring (MAM) 2006 was the third warmest in coastal southern Greenland since 1958 (Table 2, data from Cappelen et al. 2007). 2006 statistics for the southeast Greenland station Tasiilaq do not suggest abnormally warm air temperatures and may be affected by above normal spring and summer sea-ice concentration anomalies (http://nsidc.colorado.edu/data/seaice_index/).

Table 2. Year 2006 statistics relative to select Greenland land station surface air temperature observations spanning the period 1958-2006

Site			Statistic	DJF	MAM	JJA	SON	Ann.
Pituffik/Thule AFB			Anomaly [K]	-0.2	3.1	1.3	2.0	1.5
	Latitude N	76.5	Rank	25	1	5	8	10
	Longitude W	68.8	Z-score	-0.3	1.8	1.2	1.0	1.0
	Earliest Year	1961	Max. Year	1963	2006	1988	1981	2003
	Latest Year	2006	Min. Year	1984	1992	1996	1986	1992
Upernavik			Anomaly [K]	2.5	3.4	1.3	1.3	2.1
	Latitude N	72.8	Rank	18	3	7	15	8
	Longitude W	56.2	Z-score	0.4	1.7	1.1	0.7	1.1
	Earliest Year	1958	Max. Year	1963	1962	1960	1998	2003
	Latest Year	2006	Min. Year	1984	1964	1970	1958	1983
Ilulissat			Anomaly [K]	2.0	3.2	0.7	0.2	1.5
	Latitude N	69.2	Rank	23	7	12	24	12
	Longitude W	51.1	Z-score	0.3	1.1	0.6	0.0	0.7
	Earliest Year	1958	Max. Year	1963	1962	1960	1960	2003
	Latest Year	2006	Min. Year	1984	1993	1972	1986	1984
Nuuk			Anomaly [K]	2.1	2.7	0.9	0.8	1.6
	Latitude N	64.2	Rank	18	5	10	15	8
	Longitude W	51.8	Z-score	0.5	1.5	0.8	0.6	1.0
	Earliest Year	1958	Max. Year	1963	1962	2003	1960	2003
	Latest Year	2006	Min. Year	1984	1993	1972	1982	1984
Prins Christian Sund			Anomaly [K]	1.5	1.5	1.1	0.6	1.2
	Latitude N	60	Rank	9	3	6	17	4
	Longitude W	43.2	Z-score	0.8	1.3	1.4	0.4	1.2
	Earliest Year	1958	Max. Year	1979	2005	2005	2003	2005
	Latest Year	2006	Min. Year	1993	1989	1992	1982	1993
Tasiilaq			Anomaly [K]	1.8	0.4	0.6	0.0	0.7
	Latitude N	65.6	Rank	6	22	14	28	14
	Longitude W	22	Z-score	1.0	0.1	0.5	-0.2	0.5
	Earliest Year	1958	Max. Year	2003	2004	2003	2002	2003
	Latest Year	2006	Min. Year	1968	1990	1983	1971	1983
Danmarkshavn			Anomaly [K]	3.6	2.9	-0.5	1.4	1.8
	Latitude N	76.8	Rank	4	3	43	10	2
	Longitude W	18.8	Z-score	1.8	2.1	-1.0	0.8	1.9
	Earliest Year	1958	Max. Year	2005	1976	2003	2002	2005
	Latest Year	2006	Min. Year	1967	1966	1983	1971	1983

Anomalies and Z-scores are computed with respect to the 1971-2000 base period.

Extreme cold years including 1884, 1992, 1993, 1983, 1984, are associated with volcanic cooling (Box, 2002; Hanna et al. 2005). Other record-setting cold years, e.g. 1887, or 1899, may result from other factors than volcanism, such as positive sea ice concentration anomalies.

Significant increases 1958-2006 in Greenland margin summer temperatures and runoff, record-high 2003 summer temperature and 2005 snowmelt records, and a highly significant correlation of recent Greenland with Northern Hemisphere temperatures, collectively suggest that an expected signal of the GrIS to global warming may be emerging (Hanna et al. 2007). This signal now appears to be distinct from natural/regional climatic fluctuations, such as those related to changes in the North Atlantic Oscillation (e.g. Hanna & Cappelen 2003).

The new Greenland summer warmth and snowmelt records are consistent in timing with recent increased losses of summer Arctic sea ice (e.g. Comiso 2006, Richter-Menge et al. 2006). Indeed, reduced extent and duration of winter sea-ice should expose Greenland to enhanced warm air advection from surrounding seas, lengthening snowmelt and runoff seasons and possibly enhance snow accumulation, the latter a negative feedback for ice sheet response to climate warming.

Considering available station data that are continuous and begin before 1900 (Table 3), the year 2006 is not outstanding. In this longer perspective, only 2003 at Tasiilaq is outstanding in recent decades. Over the past century, years in Greenland that register as abnormally warm, 1929, 1932, 1941, 1947, and 1960 are outstanding, having temperatures warmer than observed recently. Increases in GrIS melt and runoff during this past century warm period must have been significant and were probably even larger than that of the most recent last decade (1995-2006).

Table 3. Year 2006 statistics relative to select Greenland land station surface air temperature observations that are continuous and begin before 1900

Site			Statistic	DJF	MAM	JJA	SON	Ann.
Upernavik			Anomaly [K]	2.5	3.4	1.3	1.3	2.1
	Latitude N	72.8	Rank	36	17	9	34	18
	Longitude W	56.2	Z-score	0.6	1.3	1.5	0.8	1.3
	Earliest Year	1873	Max. Year	1947	1932	1960	1928	1947
	Latest Year	2006	Min. Year	1898	1896	1922	1917	1887
Ilulissat			Anomaly [K]	2.0	3.2	0.7	0.2	1.5
	Latitude N	69.2	Rank	41	22	20	56	22
	Longitude W	51.1	Z-score	0.6	1.0	1.0	0.3	1.0
	Earliest Year	1873	Max. Year	1963	1932	1960	1960	1947
	Latest Year	2006	Min. Year	1898	1887	1972	1884	1884
Nuuk			Anomaly [K]	2.1	2.7	0.9	0.8	1.6
	Latitude N	64.2	Rank	33	16	33	31	16
	Longitude W	51.8	Z-score	0.8	1.3	0.7	0.8	1.1
	Earliest Year	1873	Max. Year	1947	1932	1948	1960	1941
	Latest Year	2006	Min. Year	1984	1993	1914	1898	1884
Tasiilaq			Anomaly [K]	1.8	0.4	0.6	0	0.7
	Latitude N	65.6	Rank	20	57	51	61	36
	Longitude W	22.0	Z-score	0.9	0.0	0.1	-0.1	0.4
	Earliest Year	1895	Max. Year	1929	1929	2003	1941	2003
	Latest Year	2006	Min. Year	1918	1990	1983	1917	1899

Anomalies and Z-scores are computed with respect to the 1971-2000 base period.

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References

Bindschadler, R., (2006), Hitting the ice where it hurts, *Science, 311*, 1720-1721.

Box, J.E., L. Yang, J. Rogers, D. Bromwich, L.-S. Bai, K. Steffen, J. Stroeve, and S.-H. Wang (2005), Extreme precipitation events over Greenland: consequences to ice sheet mass balance, American Meteorological Society, 8th International Conference on Polar Meteorology and Oceanography, 9-13 January 2005, San Diego, CA, paper 5.2.

Box, J.E., (2002), Survey of Greenland instrumental temperature records: 1873-2001, *Int. J. Climatol., 22*, 1829-1847.

Box, J.E., D.H. Bromwich, B.A. Veenhuis, L.-S. Bai, J.C. Stroeve, J.C. Rogers, K. Steffen, T. Haran, and S.-H. Wang (2006), Greenland ice sheet surface mass balance variability (1988-2004) from calibrated Polar MM5 output, *J. Clim., 19*, 2783-2800.

Cappelen J., E.V. Laursen, P.V. Jørgensen and C. Kern-Hansen, DMI monthly Climate Data Collection 1768-2006, Denmark, The Faroe Islands and Greenland, DMI Technical Report No. 07-06, Spring 2007.

Chen, J.L., C.R. Wilson, and B.D. Tapley (2006), Satellite gravity measurements confirm accelerated melting of Greenland ice sheet, *Science, 313*, 1958-1960.

Chylek, P., M.K. Dubey, and G. Lesins (2006), Greenland warming of 1920-1930 and 1995-2005, *Geophys. Res. Lett., 33*, L11707, doi:10.1029/2006GL026510.

Comiso, J.C. (2006), Arctic warming signals from satellite observations, *Weather, 61*, 70-76.

Gregory, J., and P. Huybrechts (2006), Ice-sheet contributions to future sea-level change, *Phil. Trans. Roy. Soc. A, 364*, 1709.

Gregory, J. M., P. Huybrechts, and S. C. B. Raper (2004), Climatology - Threatened loss of the Greenland ice-sheet, *Nature, 428*, 616.

Hanna, E., and J. Cappelen (2003), Recent cooling in coastal southern Greenland and relation with the North Atlantic Oscillation, *Geophys. Res. Lett., 30*, 1132, doi:10.1029/2002GL015797.

Hanna, E., P. Huybrechts, I. Janssens, J. Cappelen, K. Steffen, and A. Stephens (2005), Runoff and mass balance of the Greenland ice sheet: 1958-2003, *J. Geophys. Res., 110*, D13108, doi:10.1029/2004JD005641.

Hanna, E., J. McConnell, S. Das, J. Cappelen, and A. Stephens (2006), Observed and modeled Greenland ice sheet snow accumulation, 1958-2003, and links with regional climate forcing, *J. Clim., 19*, 344-358.

Hanna, E., P. Huybrechts, K. Steffen, J. Cappelen, R. Huff, C. Shuman, T. Irvine-Fynn, S. Wise, and M. Griffiths (2007), Increased runoff from melt from the Greenland Ice Sheet: a response to global warming, *J. Clim.*, in press.

Howat, I.M., I. Joughin, S. Tulaczyk, and S. Gogineni (2005), Rapid retreat and acceleration of Helheim Glacier, east Greenland, *Geophys. Res. Lett., 32*, L22502, doi: 10.1029/2005GL024737.

Howat, I.M., I. Joughin, and T.A. Scambos (2007), Rapid changes in ice discharge from Greenland outlet glaciers, *Science, 315*, 1559-1561.

Huybrechts, P., J. Gregory, I. Janssens, and M. Wild (2004), Modelling Antarctic and Greenland volume changes during the 20th and 21st centuries forced by GCM time slice integrations, *Glob. Planet. Change, 42*, 83-105.

IPCC (2007), Climate Change 2007: The Physical Science Basis Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Summary for Policymakers, http://www.ipcc.ch/SPM2feb07.pdf.

Janssens, I., and P. Huybrechts (2000), The treatment of meltwater retention in mass-balance parameterizations of the Greenland ice sheet, *Ann. Glaciol., 31*, 133-140.

Johannessen, O.M., et al. (2004), Arctic climate change: observed and modelled temperature and sea-ice variability, *Tellus, 56*, 328-341.

Johannessen, O. M., K. Khvorostovsky, M.W. Miles, and L.P. Bobylev (2005), Recent ice-sheet growth in the interior of Greenland, *Science, 310*, 1013-1016.

Krabill, W., W. Abdalati, E. Frederick, S. Manizade, C. Martin, J. Sonntag, R. Swift, R. Thomas, W. Wright, and J. Yungel (2000), Greenland ice sheet: High-elevation balance and peripheral thinning, *Science, 289*, 428-430.

Krabill, W., et al. (2004), Greenland ice sheet: Increased coastal thinning, *Geophys. Res. Lett., 31*, L24402, doi:10.1029/2004GL021533.

Liang S., J. Stroeve, and J.E. Box (2005), Mapping daily snow/ice shortwave broadband albedo from Moderate Resolution Imaging Spectroradiometer (MODIS): The improved direct retrieval algorithm and validation with Greenland in situ measurement, *J. Geophys. Res., 110*, D10109, doi:10.1029/2004JD005493.

Luthcke, S.B., H.J. Zwally, W. Abdalati, D.D. Rowlands, R.D. Ray, R.S. Nerem, F.G. Lemoine, J.J. McCarthy, and D.S. Chinn (2006), Recent Greenland ice mass loss by drainage system from satellite gravity observations, *Science, 314*, 1286-1289.

Murray, T. (2006), Climate change: Greenland's ice on the scales, *Nature, 443*, 277-278.

Parizek, B.R., and R.B. Alley (2004), Implications of increased Greenland surface melt under global-warming scenarios: ice-sheet simulations, *Quat. Sci. Rev., 23*, 1013.

Ramillien, G., A. Lombard, A. Cazenave, E.R. Ivins, M. Llubes, F. Remy, and R. Biancale (2006), Interannual variations of the mass balance of the Antarctic and Greenland ice sheets from GRACE, *Glob. Planet. Change, 53*, 198-208.

Richter-Menge, J., and others (2006), Arctic, in State of the Climate in 2005, Shein, K.A., ed., *B. Am. Meteorol. Soc., 87*, S46-S53.

Rignot, E., and P. Kanagaratnam (2006), Changes in the velocity structure of the Greenland ice sheet, *Science, 311*, 986-990, doi:10.1029/2002GL015797.

Shepherd, A., and D. Wingham (2007), Recent sea-level; contributions of the Antarctic and Greenland ice sheets, *Science, 315*, 1529-1532.

Stearns, L.A., and G.S. Hamilton (2007), Rapid volume loss from two East Greenland outlet glaciers quantified using repeat stereo satellite imagery, *Geophys. Res. Lett., 34*, L05503, doi:10.1029/2006GL028982.

Thomas, R., et al. (2003), Investigation of surface melting and dynamic thinning on Jakobshavn Isbrae, Greenland, *J. Glaciol., 49*, 231-239.

Thomas, R., E. Frederick, W. Krabill, S. Manizade, C. Martin, and A. Mason (2005), Elevation changes on the Greenland ice sheet from comparison of aircraft and ICESat laser-altimeter data, *Ann. Glaciol., 42*, 77-82.

Thomas, R., E. Frederick, W. Krabill, S. Manizade, and C. Martin (2006), Progressive increase in ice loss from Greenland, *Geophys. Res. Lett.*, art. no. L10503.

Velicogna, I., and J. Wahr (2006), Acceleration of Greenland ice mass loss in spring 2004, *Nature, 443*, 329-331.

Zwally, H.J., W. Abdalati, T. Herring, K. Larson, J. Saba, and K. Steffen (2002), Surface melt-induced acceleration of Greenland ice-sheet flow, *Science, 297*, 218-222.

Zwally et al. (2005), Mass changes of the Greenland and Antarctic ice sheets and shelves and contributions to sea-level rise: 1992-2002, *J. Glaciol. 51*, 509-527.

Recent Changes in Vegetation

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Vegetative Boundaries

In northern latitudes, there is a series of vegetation and plant life form boundaries that are associated with temperature (Figure 1). They generally move northwards as climate becomes warmer although advance is not expected to be uniform, in part due to differences in individual species response. The latitudinal treeline is the most obvious one, and is associated with the isotherm for July mean monthly temperature of about 11° C. North of this boundary, one finds the shrubline, and further north, the tundra becomes completely devoid of woody plants. Along this latitudinal gradient, the height of the vegetation decreases, the complexity of the plant canopy is reduced and the biodiversity generally decreases. There also tends to be a decrease in the carbon annually captured by photosynthesis and an increase in albedo (i.e. the incoming solar radiation reflected from the land surface). As the latitudinal distribution of plant life forms are associated with climate, vegetation zone dynamics and shifts in their composition are considered one of the major processes that will respond to a warming Arctic (Callaghan et al., 2004a).

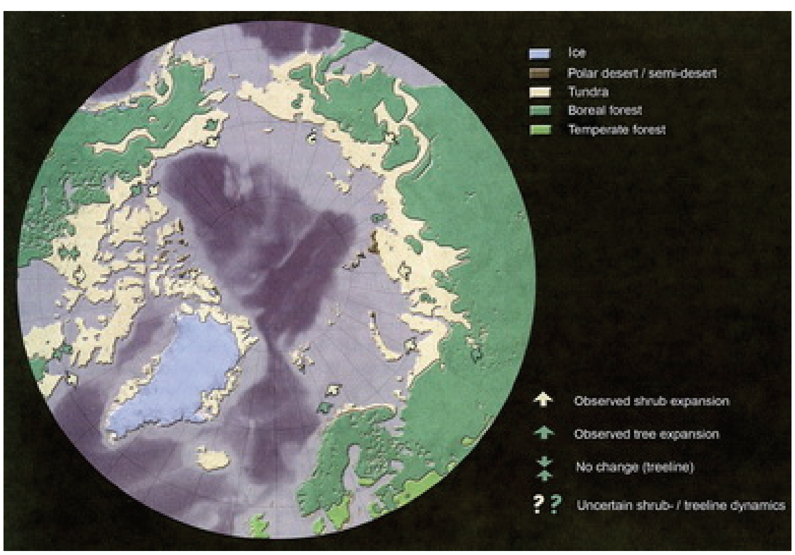

Figure 1. Observed arctic vegetation dynamics (modified from Kaplan et al., 2003 and Callaghan et al., 2005).

Recent vegetation dynamics observations across the Arctic show that, in general, shrubs have become more abundant and taller. A study in northern Alaska (Tape et al., 2006) showed that both larger and smaller shrub species have increased in size, abundance and extent over the last 50 years. As well as increasing in size and filling in empty patches, the shrubs were colonizing new areas (Figure 2). Results do vary regionally. For instance, Shvartsman et al. (1999) pointed to a decrease in shrubs along the Pechora River in western Russia (66.1° N, 57.1° E) between 1960 and 1983, a change they attributed to an increase in trees. Preliminary results from Lantz & Henry (unpublished data, Tape et al., 2006) showed a recent expansion of the shrub cover on the Canadian Mackenzie River delta (69.1° N, 135.1° W). In the Northwest Territories (65.1° N, 111.5° W), P. Grogan (personal communication, 2005; Tape et al., 2006) observed an increase in shrubs on floodplains and stream channels, whereas in Labrador (58.1° N, 72.1° W), Payette (2006) found that alder had increased in conjunction with a northward migration of the treeline.

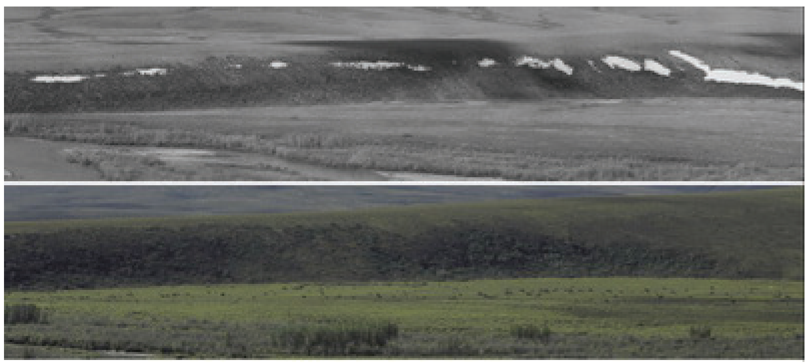

Figure 2. Large shrubs have colonized a river terrace that was virtually free of large shrubs in 1949. The new shrubs are more than 2 m high. In the foreground are poplar trees. Photo from the Chandler River located at 68° 25.14' N, 161° 15.24' W: 7/4/1948 and 7/29/2001. (Tape et al., 2006).

Other evidence of widespread change of vegetation in the tundra regions come from trends in tundra greenness as detected by satellites. The Normalized Difference Vegetation Index (NDVI) is a measure of greenness derived from reflectance of the surface in the red and near-infrared channels. If the climate warms, higher NDVI values might be expected to shift northward. Earlier global studies of NDVI changes indicated a general pattern of increased NDVI in the region between 40-70° N during the period 1981-1999 (Myneni et al., 1997, 2001; Zhou et al., 2001; Lucht et al., 2002). Stow et al. (2004) noted the largest NDVI increase for the period 1982-1999 in northern Russia (north of 65° N, between 70 and 140° E), the North Slope of Alaska, and parts of northern Canada and Scandinavia. Studies of the NDVI in the tundra area of northern Alaska indicate an increase of 17% in NDVI values in this region. A follow-up to this study shows linear trends in Arctic tundra vegetation greenness over the period 1982-2005 as positive over Eurasia (Yamal) and North America, as observed with NOAA AVHRR satellite. Although general trends of greening were observed, there are different magnitudes between the tundra biome in Eurasia and North America. For instance, in the region south of 70 degrees north, the rate of change is +0.58%/yr over the North American Arctic compared to +0.34%/yr over the Yamal Arctic.

More recent studies based on the 1981-2005 AVHRR record show that there is not a simple linear positive trend in NDVI across all of the North corresponding to warmer temperatures (Goetz et al., 2005, Bunn et al., 2007). Eighty-eight percent of the northern high latitudes show no significant deterministic trend in satellite-derived NDVI values. Nine percent of the areas show increased NDVI, while 3% indicate a decline. In the forested areas, 6% of the areas show a decline while 4% indicate an increase; whereas, in the tundra and other shrubby areas, 1%

show a decline, while 6% reveal an increase in NDVI. The increasing NDVI values in the tundra areas is thought to be primarily caused by increasing density of shrubs, while the decline of NDVI in the forested areas is attributed to increased moisture stress (Bunn et al., 2007).

Treeline dynamics are also complex. Although most arctic areas report a recent advance of this vegetation boundary, there are considerable inertia effects in some areas. For example, a broad-scale study by Lloyd (2005) revealed a treeline advance occurring throughout three separate regions of Alaska. Although treeline advance was detected at all locations, the timing of the advance varied by more than a century among regions. Besides timing, the spatial scale of the advance also varied among sites.

Several large scale studies in northern and eastern Canada making use of both macro-fossil tree remains and remote sensing data, revealed a long-term stability of the treeline ecotone during the past 2000-3000 years (Lavoie & Payette, 1996; Masek, 2001; Payette, 2006) (Figure 1). Recently, the treelines in the forest-tundra areas of Quebec have risen slightly either through establishment of seed-origin white spruce Picea glauca or through height growth of stunted spruce already established on the tundra hilltops. However, invasion of tundra and alpine areas by trees occurred only gradually and seems seriously hampered by local topographic factors as well as by harsh wind-exposure conditions (Gamache & Payette, 2005; Payette, 2006; Caccianiga & Payette, 2006).

In western Canada, which is experiencing a similar warming and prolonging of the growing season as Alaska (Chapin et al., 2005; Stafford et al., 2000), white spruce is currently invading the southern exposed alpine areas, but north-facing slopes have only experienced a densification (+40 - 65%), (Danby & Hik, 2007), fig.1.

Recent-most investigations in northern Europe report a significant sprouting of seedlings and saplings high above the treeline (Kullman, 2002; Löffler et al., 2004; Dalen & Hofgaard, 2005; Kullman & Kjällgren, 2006; Truong et al., 2007; Van Bogaert, unpublished data) as well as important treeline ecotone densification processes (Tømmervik et al., 2004; Hållmarker, 2002; Figure 1). In the southern Swedish Scandes Mountains (63.1° N, 12.2° E), the mountain birch treeline (Betula pubescens ssp. tortuosa) rose on average by about 75 m in elevation during the last century (Kullman, 2001, 2002, 2003), while in northernmost Europe (Sweden) the altitudinal increase rate of the treeline is estimated on 0.5 m yr^{-1} or 40 m per °C summer temperature increase (Callaghan et al., 2004b; Figure 3). However, Dalen and Hofgaard (2005) studied recent treeline conditions across northern Europe and came to the conclusion that regional differentiation needs to be considered. In the southern and northern Scandes mountains (Dovre 62.2° N, 9.4° E; Abisko 68.3° N, 18.9° E resp.) the treeline was in a rejuvenating and possibly expanding state, while in northernmost Europe (Finnmark, Norway 69.7° N, 24.0° E) a receding treeline was observed. Most likely the latter observation is related to a decrease in the length of the growing season due to higher winter precipitation (Høgda et al., 2001) and a higher number of reindeer (Ims and Kosmo, 2001).

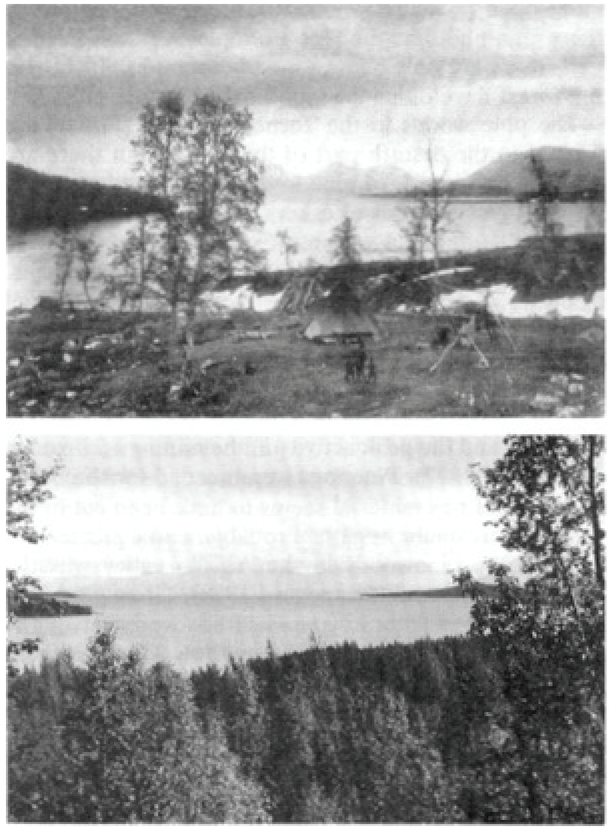

Figure 3. Forest growth in northern Sweden between 1906 (top) and 1986 (Emanuelsson, 1987). Attribution is difficult: temperature has increased but land use has also changed. The Sami camp in the early 20th century no longer exists and the intensity of land use has decreased.

During the past century, considerable treeline shifts have been reported from the Russian sub-arctic. Both southern and Polar Ural mountain areas have noted a rising treeline within the past century: 60 to 80 m, and 20 to 40 m respectively (Moiseev & Shiyatov, 2001, 2003; Shiyatov et al., 2005). The forested area in northeastern European Russia (Katenin, 2004; personal communication, 2007) and the world's northernmost forest range of Ary-Mas in northwest Siberia have also expanded displacing tundra patches at a rate of 3-10 m per year (Kharuk et al., 2006). Even though data from Russia are still rather scarce and inaccessible (Figure 1), there is also knowledge about contrasting dynamics. For instance, Lapenis et al. (2005) found that during the past 50 years the fraction of leaves and needles has decreased in the vast northern taiga zone of Siberia, where the climate has become warmer, but drier. Additionally, Russia is rather densely populated at the tundra taiga transition zone, causing some large-scale southward treeline recessions due to human activities (Vlassova, 2002). Therefore, attribution is complex and uncertain, and is likely to change from location to location.

Fire and insect outbreaks are inherently connected to the boreal forest and changes in these regimes may alter forest type, structure and extension dramatically (i.e. McCullough et al., 1998; Stocks et al., 1998; Karlsson et al., 2004; Tenow et al., 2004; Juday et al., 2005). Additionally, since the 1950s, growth responses of treeline trees to climate seem to diverge more and more (Lloyd & Fastie, 2002; Wilmking et al., 2004, 2005; Driscoll et al., 2006; Pisaric et al., 2006) challenging both our understanding of climate change responses and climate reconstructions.

References

Bunn, A.G., S.J. Goetz, J.S. Kimball, and K. Zhang (2007) Northern high-latitude ecosystems respond to climate change. *EOS, 88*(34): 333-340.

Caccianiga, M., and S. Payette (2006) Recent advance of white spruce (Picea glauca) in the coastal tundra of the eastern shore of Hudson Bay (Québec, Canada). *Journal of Biogeography, 33*, 12, 2120-2135.

Callaghan T.V., L.O. Björn, Y. Chernov, T. Chapin, T.R. Christensen, B. Huntley, R.A. Ims, M. Johansson, D. Jolly, S. Jonasson, N. Matveyeva, N. Panikov, W. Oechel, G. Shaver, S. Schaphoff, and S. Sitch (2004a) Effects of changes in climate on landscape and regional processes, and feedbacks to the climate system. *Ambio, 33*, 7, 459-468.

Callaghan T.V., M. Johansson, O.W. Heal, N.R. Saelthun, L.J. Barkved, N. Bayfield, O. Brandt, R. Brooker, H.H. Christiansen, M. Forchhammer, T.T. Høye, O. Humlum, A. Järvinen, C. Jonasson, J. Kohler, B. Magnusson, H. Meltofte, L. Mortensen, S. Neuvonen, I. Pearce, M. Rasch, L. Turner, B. Hasholt, E. Huhta, E. Leskinen, N Nielsen, and P. Siikamäki (2004b) Environmental changes in the North Atlantic region: SCANNET as a collaborative approach for documenting, understanding and predicting changes. *Ambio, 13*, 39-50.

Callaghan T.V., L.O. Björn, F.S. Chapin III, Y. Chernov, T.R. Christensen, B. Huntley, R. Ims, M. Johansson, D.J. Riedlinger, S. Jonasson, N. Matveyeva, W. Oechel, N. Panikov, and G. Shaver (2005) Arctic Tundra and Polar Desert Ecosystems. Chapter 7, In: *Arctic Climate Impact Assessment*. Arctic Council. Cambridge University, 243-352.

Chapin, F.S., M. Sturm, M.C. Serreze, J.P. McFadden, J.R. Key, A.H. Lloyd, A.D. McGuire, T.S. Rupp, A.H. Lynch, J.P. Schimel, J. Beringer, W.L. Chapman, H.E. Epstein, E.S. Euskirchen, L.D. Hinzman, G. Jia, C.L. Ping, K.D. Tape, C.D.C. Thompson, D.A. Walker, and J.M. Welker (2005) Role of land-surface changes in Arctic summer warming. *Science, 310*:657-660.

Dalen, L., and A. Hofgaard (2005). Differential regional treeline dynamics in the Scandes Mountains. *Arctic, Antarctic, and Alpine Research, 37*, 3, 284-296.

Danby, R.K., and D.S. Hik (2007) Variability, contingency and rapid change in recent subarctic alpine tree line dynamics. *Journal of Ecology, 95*(2), 352-363.

Driscoll, W., G. Wiles, R. D'Arrigo, and M. Wilmking (2006) Divergent tree growth response to recent climatic warming, Lake Clark National Park and Preserve, Alaska. *Geophysical Research Letters, 32*: 2020.

Emanuelsson, U. (1987) Human influence on vegetation in the Torneträsk area during the last three centuries. *Ecological Bulletins, 38*, 95-111.

Gamache, I., and S. Payette (2005) Latitudinal response of subarctic tree lines to recent climate change in eastern Canada. *Journal of Biogeography, 32*(5).

Goetz, S.J., A.G. Bunn, G.J. Friske, and R.A. Houghton (2005) Satellite-observed photosynthetic trends across boreal North America associated with climate and fire disturbance. *Proceedings of the National Academy of Sciences, 102*, 13521-13525.

Hållmarker, M. (2002) Skogsgränsförändringar i Abisko. Master thesis in Environmental Science. Götenborg University, 37 p.

Høgda, K.A., S.R. Karlsen, and I. Solheim (2001) Climatic change impact on growing season in Fennoscandia studied by a time series of NOAA AVHRR NDVI data. *Proceedings of IGARSS 2001*. Sydney, Australia 9-13 July 2001:1338-1340.

Ims, A., and A.J. Kosmo (2001) Høyeste reintall for distriktene i Vest-Finnmark. Alta: reindriftsforvaltningen, 153 p.

Juday G.P., V. Barber, P. Duffy, H. Linderholm, S. Rupp, S. Sparrow, E. Vaganov, and J. Yarie (2005) Forests, Land Management and Agriculture. Chapter 14 In: *Arctic Climate Impact Assessment*. Arctic Council. Cambridge University, 781-862.

Kaplan J.O., N.H. Bigelow, I.C. Prentice, S.P. Harrison, P.J. Bartlein, T.R. Christensen, W. Cramer, N.V. Matveyeva, A.D. McGuire, D.F. Murray, V.Y. Razzhivin, B. Smith, D.A. Walker, P.M. Anderson, A.A. Andreev, L.B. Brubaker, M.E. Edwards, and A.V. Lozhkin (2003) Climate change and Arctic ecosystems II. Modeling, paleodata-model comparisons, and future projections. *Journal of Geophysical Research, 108*, 8171.

Karlsson S., H. Bylund, and O. Tenow (2004) Effects of outbreaks of the autumnal moth Epirrita autumnata on the mountain birch forest. *Svensk Botanisk Tidskrift, 98*, 3-4, 162-172.

Kateninen, A.E. (2004) Structure of the plant communities of spruce open woodland of different ages on the taiga-tundra boundary in the east-European part of Russia. ACIA International Scientific Symposium on Climate Change in the Arctic, Reykjavik, Iceland, November 9-12, 2004.

Katenin, A.E. (2007) Personal communication, Russia, May 2007.

Kharuk, V.I., K.J. Ranson, S.T. Im, and M.M. Naurzbaev (2006) Forest-tundra larch forests and climatic trends. *Russian Journal of Ecology, 37*, 5.

Kullman, L. (2001) 20th century climate warming and tree-limit rise in the southern Scandes of Sweden. *Ambio, 30*, 72- 80.

Kullman, L. (2002) Rapid recent range-margin rise of tree and shrub species in the Swedish Scandes. *Journal of Ecology, 9*0, 68-77.

Kullman, L. (2003) Recent reversal of Neoglacial climate cooling trend in the Swedish Scandes as evidenced by mountain birch treelimit rise. *Global and Planetary Change, 36*, 77-88.

Kullman, L., and L. Kjällgren (2006) Holocene pine tree-line evolution in the Swedish Scandes: recent treeline rise and climate change in a long-term perspective. *Boreas, 35*, 159-168.

Lapenis, A., A. Shvidenko, D Shepaschenko, S. Nilsson, and A. Aiyyer (2005) Acclimation of Russian forests to recent changes in climate. *Global Change Biology, 11*, 2090-2102.

Lavoie, C., and S. Payette (1996) The long-term stability of the boreal forest limit in subarctic Quebec. *Ecology, 77*, 4, 1226-1233.

Lloyd, A. (2005) Ecological histories from Alaskan treelines provide insight into future change. *Ecology, 86*, 77, 1687-1695.

Lloyd, A.H., and C.L. Fastie (2002) Spatial and temporal variability in the growth and climate response of treeline trees in Alaska. *Climatic Change, 52*, 4, 481-509.

Löffler, J., A. Lundberg, O. Rössler, A. Brauning, G. Jung, R. Pape, and D. Wundram (2004) The alpine treeline under changing land use and changing climate: Approach and preliminary results from continental Norway. *Norsk Geografisk Tidsskrift-Norwegian Journal of Geography, 58*, 183-193.

Lucht W., I.C. Prentice, R.B. Myneni, S. Sitch, P. Friedlingstein, W. Cramer, P. Bousquet, W. Buermann, and B. Smith (2002) Climatic control of the high-latitude vegetation greening trend and Pinatubo effect. *Science, 296*, 5573, 1687-1689.

Masek, J.G. (2001) Stability of boreal forest stands during recent climate change: evidence from Landsat satellite imagery. *Journal of Biogeography, 28*, 8, 967-976.

McCullough D.G., R.A. Werner, and D. Neumann (1998) Fire and insects in northern and boreal forest ecosystems of North America. *Annual Review of Entomology, 43*, 107-127.

Moiseev, P., and S. Shiyatov (2001) Vegetation dynamics at the treeline ecotone in the Ural highlands, Russia. *Alpine Biodiversity in Europe, S.*, 423-435.

Moiseev, P.A., and S.G. Shiyatov (2003) Vegetation dynamics at the tree line ecotone in the Ural Highlands, Russia. In: *Alpine Biodiversity in Europe* (eds. L. Nagy, G. Grabherr, C. Körner, and D.B.A. Thompson), Springer, Berlin, 423-435.

Myneni R.B., C.D. Keeling, C.J. Tucker, G. Asrar, and R.R. Nemani (1997) Increased plant growth in the northern high latitudes from 1981 to 1991. *Nature, 38*, 698-702.

Myneni R.B., J. Dong, C.J. Tucker, R.K. Kaufmann, P.E. Kauppi, J. Liski, L. Zhou, V. Alexeyev, and M.K. Hughes (2001) A large carbon sink in the woody biomass of Northern forests. *PNAS, 98*, 26, 14784-14789.

Payette, S. (2006) Contrasted dynamics of northern Labrador tree lines caused by climate change and migrational lag. *Ecology, 88*, 3, 770-780.

Pisaric, M.F., S. Carey, S.V. Kokelj, and D. Youngblut (2006) Anomalous 20th Century Tree Growth, Mackenzie Delta, Northwest Territories, Canada. *Eos Trans. AGU, 87*, (52), Geophysical Union, Fall Meeting 2006.

Shiyatov, S.G., M.M. Terent'ev, and V.V. Fomin (2005) Spatiotemporal dynamics of forest-tundra communities in the Polar Urals. *Russian Journal of Ecology, 36*, 2, 69-75.

Shvartsman Y.G., V.M. Barzut, S.V. Vidyakina, and S.A. Iglovsky (1999) Climate variations and dynamic ecosystems of the Arkhangelsk. *Chemosphere - Global Change Science, 1*, 417-428.

Stafford, J.M., G. Wendler, and J. Curtis (2000) Temperature and precipitation of Alaska: 50 year trend analysis. *Theoretical and Applied Climatology, 67*, 1-2, 33-44.

Stocks B.J., M.A. Fosberg, T.J. Lynham, L. Mearns, B.M. Wotton, O. Yang, J.-Z. Jin, K. Lawrence, G.R. Hartley, J.A. Mason, and D.W. McKenney (1998) Climate Change and Forest Fire Potential in Russian and Canadian Boreal Forests. *Climatic Change, 38*, 1, 1-13.

Stow D.A., A. Hope, D. McGuire, D. Verbyla, J. Gamon, F. Huemmrich, S. Houston, C. Racine, M. Sturm, and K. Tape (2004) Remote sensing of vegetation and land-cover change in Arctic Tundra Ecosystems. *Remote Sensing of Environment, 89*:281- 308.

Tape, K., M. Sturm, and C. Racine (2006) The evidence for shrub expansion in Northern Alaska and the Pan-Arctic. *Global Change Biology, 12*, 686-702.

Tenow O., H. Bylund, P.S. Karlsson, and J. Hoogesteger (2004) Rejuvenation of a mountain birch forest by an Epirrita autumnata (Lepidoptera: Geometridae) outbreak. *Acta Oecologica, 25*, 1-2, 43-52.

Truong, C., A.E. Palmé, and F. Felber (2007) Recent invasion of the mountain birch Betula pubescens ssp. tortuosa above the treeline due to climate change: genetic and ecological study in northern Sweden. *Journal of Evolutionary Biology, 20*, 1, 369-380.

Tømmervik, H., B. Johansen, I. Tombre, D. Thannheiser, K.A. Hogda, E. Gaare, and F.E. Wielgolaski (2004) Vegetation changes in the Nordic mountain birch forest: the influence of grazing and climate change. *Arctic, Antarctic, and Alpine Research, 36*, 3, 323-332.

Vlassova, T.K. (2002) Human impacts on the tundra-taiga zone dynamics: the case of the Russian lesotundra. *Ambio, Special Report, 12*, 30-6.

Wilmking, M., G.P. Juday, V.A. Barber, and H.S.J. Zald (2004) Recent climate warming forces contrasting growth responses of white spruce at treeline in Alaska through temperature thresholds. *Global Change Biology, 10*(10), 1724-1736.

Wilmking, M., R. D'Arrigo, G. Jacoby, and G. Juday (2005) Increased temperature sensitivity and divergent growth trends in circumpolar boreal forests. *Geophysical Research Letters, 32*, 15, L15715.1-L15715.4.

Zhou L., C.J. Tucker, R.K. Kaufmann, D. Slayback, N.V. Shabanov, I. Fung, and R.B. Myneni (2001) Variations in northern vegetation activity inferred from satellite data of vegetation index during 1981 to 1999. *Journal of Geophysical Research, 106*, D17, 20,069-20,083.

State of Reindeer Herds

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Reindeer are also known as caribou. Rangifer herds across the circumpolar north have long been characterized by periods of abundance and periods of scarcity. Recent population estimates indicate we may be entering a period of declining numbers. Populations that have been increasing at a steady rate since the early to mid 1970's are either showing signs of peaking or beginning to decline. Figure 1 shows the current status of the main migratory herds across the circumpolar north.

- The largest herds Western Arctic (490,000 in 2003), the Leaf River/George River Herds (1,000,000 in 2004) and the Taimyr herd in Russia (600,000 in 2003) have slowed from previously high growth rates or remain stable between the last 2 consecutive counts.
- The Porcupine Caribou Herd was the first herd to decline, dropping from a high of 178,000 in 1989 to 123,000 in 2001.
- In the central barrens of Northwest Territories and Nunavut herds have declined by as much as 80% in the last 5 years. These dramatic declines have local agencies and user communities preparing for resource shortages in the near future.
- See: http://www.nwtwildlife.com/NWTwildlife/caribou/newsreleasesept06.htm, and (http://www.arcticpeoples.org/2007/02/03/canadas-disappearing-caribou/)
- While herds in central Russia have declined, the Chukotka herd in eastern Siberia has increased greatly since declines in the domestic reindeer industry following the collapse of the Soviet Union. There is some speculation that domestic stock may have augmented wild reindeer population in the region.

Although many predicted that herds would not continue to expand, the increased threats of climate change, increased industrial expansion in the north and the increased sophistication and mobility of harvesters will require more careful monitoring and analysis of population response. The CircumArctic Rangifer Monitoring and Assessment (CARMA) Network (http://www.rangifer.net/carma/) was formed in response for a need to cooperate and coordinate monitoring efforts across the north. The Network will take advantage of the International Polar Year initiative to increase its monitoring and assessment activities over the next 4 years.

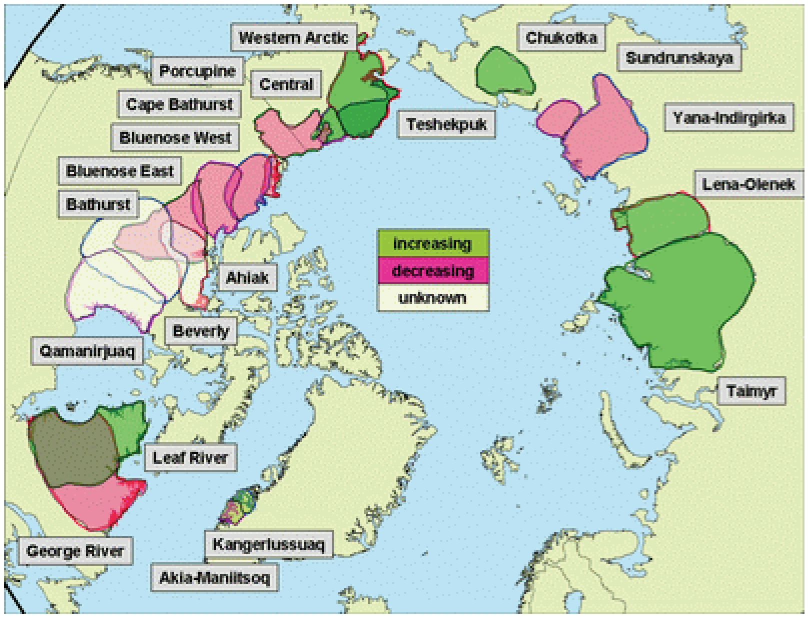

Figure 1. Current status of the main migratory herds across the circumpolar north.

Additional Information:

- Ulvevadet, B. and K. Klokov (editors) 2004. Family-based reindeer herding and hunting economies, and the status and management of wild reindeer/caribou populations. Arctic Council Report. Center for Saami Studies, University of Tromso, Norway. 170 pp.
- Russell, D., G. Kofinas, and B. Griffith. 2000. Need and opportunity for a North American caribou knowledge cooperative. *Polar Research, 19*(1): 117-130.

Goose Populations

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Goose populations are intensively monitored. Population estimates are based on simultaneous counts in wintering areas, often supplemented with data on nesting densities, ring recoveries and sightings of colour-marked individuals. Wetlands International (www.wetlands.org) is the organisation which compiles all population data with help of its Goose Specialist Group (www.geese.nl/gsg).

Geese are common in many parts of the Arctic. All Arctic populations are migratory and their annual migration routes and stop over places involve a large proportion of the Northern Hemisphere, including almost all countries in North America, Europe and North, Central and East Asia. Goose populations have a direct and significant influence on Arctic ecosystems as exemplified by recent impacts on tundra vegetation due to expanding populations and via the role played by goslings and eggs as a food source for predators in the Arctic.

Since the 1970's, many goose populations have gone through an impressive increase in size. In the last decade, the global goose population almost doubled from 12.5 million birds (Madsen et al. 1996) to a current total of 21.4 million (Wetlands International, 2006). Most of these population increases have coincided with large range extensions within the Arctic, but also into temperate regions. Changing agricultural practices have resulted in new, abundant and high quality food sources for wintering geese (Van Eerden et al. 1996, Fox et al. 2005). This has occurred while hunting pressure has decreased through improved legislative protection, a decline in the ratio of hunters per 1000 geese and the establishment of refuge areas.

The most recent review of water bird populations (Wetlands International, 2006) considers several Arctic goose populations as declining. The declines are widely distributed across all flyways indicating a possible link to phenomena acting on a circumpolar scale. Figure 1 depicts the overall distribution of trends within Arctic goose populations. For nine percent of the population, there is no or insufficient information on trends. Thirty-six percent of the populations are still increasing, thirty-two percent are stable, but twenty-three percent are declining – a proportion slightly higher than compared with ten years ago (Madsen et al. 1996).

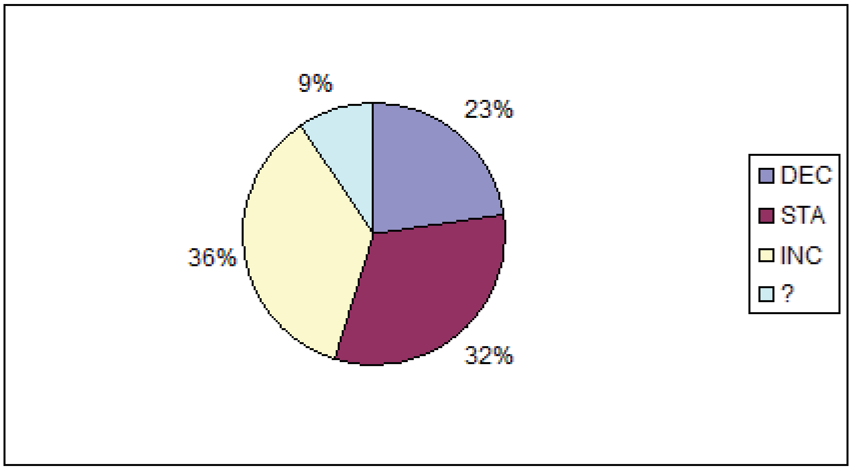

Figure 1. Trends in 47 Arctic Geese populations (Wetlands International, 2006). DEC - population decreasing; STA - population stable; INC - population increasing; ? - unknown

References

Fox, A.D., Madsen, J., Boyd, H., Kuijken, E., Norris, D.W., Tombre, I.M., and Stroud, D.A. (2005) Effects of agricultural change on abundance, fitness components and distribution of two arctic-nesting goose populations. *Glob. Change Biol., 11*: 881-893.

Madsen, J., Reed, A., and Andre, A.(1996) Status and trends of geese (Branta and Anser sp.) in the world: Review, updating and evaluation. *Gibier Faune Sauvage, Game Wildlife., 13*: 337-353.

Van Eerden, M.R., Zijlstra, M., and Van Roomen, M. (1996) The response of Anatidae to changes in agricultural practice: long term shifts in the carrying capacity for wintering waterfowl. *Gibier Faune Sauvage, Game Wildlife, 13*: 681-706.

Wetlands International (2006) Waterbirds Population Estimates - Fourth edition.

More information

Fox, A.D., Stroud, D.A., Walsh, A.J., Wilson, H.J., Norris, D.W., and Francis, I.S. (2006) The rise and fall of the Greenland White-fronted Goose: a case study in international conservation. *Brit. Birds, 99*: 242-261.

Kerbes, R.H., Kotanen, P.M., and Jefferies, R.L. (1990) Destruction of wetlands habitats by Lesser Snow Geese: a keystone species on the west coast of Hudson Bay. *J. Appl. Ecol., 27*: 242-258.

O'Connell, M.J., Huiskes, A.H.L., Loonen, M.J.J.E., Madsen, J., Klaassen, M., and Rounsevell, M. (2006) Developing an integrated approach to understanding the effects of climate change and other environmental alterations at a flyway level. Waterbirds around the World. Eds. G.C. Boere, C.A. Galbraith & D.A. Stroud. The Stationary Office, Edinburgh, UK. Pp. 385-397.

About the Report Card

The Arctic Report Card 2007 is introduced as a means of presenting clear, reliable and concise information on recent observations of environmental conditions in the Arctic, relative to historical time series records. It provides a method of updating and expanding the content of the State of the Arctic Report, published in fall 2006, to reflect current conditions.

Material presented in the Report Card is prepared by an international team of scientists and is peer-reviewed by topical experts nominated by the US Polar Research Board. The audience for the Arctic Report Card is wide, including scientists, students, teachers, decision makers and the general public interested in Arctic environment and science. The web-based format will facilitate future timely updates of the content.

Citation

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Support provided by the NOAA Climate Program Office and Arctic Research Program. The Atmosphere image is from NOAA's North Pole Web Cam, the Sea Ice photograph is provided courtesy of Jeremy Harbeck, the Biology photograph of snow geese is from Saskatchewan Schools, Canada, the Ocean photograph of oceanographic instrumentation is from the NOAA Ocean Explorer program, the Greenland image is from NASA/SVS, and the Land/Permafrost photograph is from Vladimir E. Romanovsky. Jackie Richter-Menge is the chief editor. The web page has been coordinated and developed by James E. Overland, Nancy N. Soreide, and Tracey Nakamura.